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Vol
6
1959

Contents of Volume 6

NUMBER 1

Page

M. N. SOKOLOVA : On the distribution of deep-water bottom animals in relation to their feeding habits and the character of sedimentation .. 1

CHARLES W. THOMAS : Lithology and zoology of an Antarctic bottom core .. 5

TAKASHI ICHIYE : On long waves in a stratified, equatorial ocean caused by a travelling disturbance 16

GRETHE RYTTER HASLE : A quantitative study of phytoplankton from the equatorial Pacific 39

Shorter Communications :

DAVIS A. MCGILL, NATHANIEL CORWIN and BOSTWICK H. KETCHUM : The temperature correction of oceanic inorganic phosphorus analyses 60

Letters to the Editors :

J. H. RYTHER and R. R. L. GUILLARD : Enrichment experiments as a means of studying nutrients limiting to phytoplankton production 65

JOHN H. STEELE : Observations of deep water overflow across the Iceland–Faroe Ridge 69

C. S. YENTSCH and J. H. RYTHER : Absorption curves of acetone extracts of deep water particulate matter 72

Book Review :

" Marine Ecology," by HILARY B. MOORE (reviewed by H. Brattström) 75

NUMBER 2

A. P. LISITZIN and A. V. ZHIVAGO : Marine geological work of the Soviet Antarctic Expedition, 1955–1957 77

A. E. KRISS : Micro-organisms as indicators of hydrological pnenomena in sees and oceans—I 88

TORBEN WOLLF : The hadal community, an introduction 95

JOHN A. MCGOWAN : The relationship of the distribution of the planktonic worm, Poeobius meseres Heath, to the water masses of the North Pacific 125

HENRY STOMMEL and A. B. ARONS : On the abyssal circulation of the world ocean—I. Stationary planetary flow patterns on a sphere 140

J. C. SWALLOW and B. V. HAMON : Some measurements of deep currents in the Eastern North Atlantic 155

Instrumental Notes :

S. G. NORDSTROM and T. R. FOLSOM : Suggestions for eliminating effects on protected reversing thermometers 169

J. B. HERSEY : Acoustically monitored bottom coring 170

NUMBER 3

A. E. KRISS, M. N. LEBEDEVA and I. N. MITZKEVITCH : Micro-organisms as indicators of hydrological phenomena in seas and oceans—II 173

P. J. D. HAYTER : The Ganges and Indus submarine canyons 184

A. P. JOUSÉ : Les diatommées des dépôts de fond le da partie nord-ouest de l'Océan Pacifique 187

M. N. HILL : A median valley of the Mid-Atlantic Ridge 193

A. de C. BAKER : Observations of squid at the surface in the North East Atlantic 206

H. LACOMBE : Quelques traits remarquables du relief sous-marin en Mer Mediterranée et en Mer Noire 211

HENRY STOMMEL and A. B. ARONS : On the abyssal circulation of the world ocean—II. An idealized model of the circulation pattern and amplitude in oceanic basins 217

Shorter Communications :

GILES W. MEAD : Hermaphroditism in archibenthic and pelagic fishes of the order Iniomi 234

NUMBER 3 *Continued* Page

Instrumental Note :
WILLIAM S. RICHARDSON and CHARLES J. HUBBARD : The contouring temperature recorder 239
Letters to the Editors :
GEORGE WÜST : Proposed International Indian Ocean Oceanographic Expedition, 1962–1963 245
MARGARET K. ROBINSON : Indian Ocean vertical temperature sections 249
Book Reviews : 259

NUMBER 4

THE CROMWELL CURRENT :
WARREN S. WOOSTER : Introduction – Investigations of equatorial undercurrents 263
JOHN A. KNAUSS : Measurements of the Cromwell Current 265
ROBERT S. ARTHUR : A review of the calculation of ocean currents at the equator 287
HENRY STOMMEL : Wind-drift near the Equator 298
JULES G. CHARNEY : Non-linear theory of a wind-driven homogeneous layer near the Equator 303
ALLAN R. ROBINSON : The general thermal circulation in equatorial regions 311
G. VERONIS : An approximate theoretical analysis of the equatorial undercurrent 318
GERHARD NEUMANN : Evidence for an equatorial undercurrent in the Atlantic Ocean 328
A. E. KRISS, S. S. ABYZOV and I. N. MITZKEVITCH: Micro-organisms as indicators of hydrological phenomena in seas and oceans—III 335
W. DANSGAARD : The content of heavy oxygen isotope in the water masses of the Phillipine Trench 346
D. W. MENZEL and J. H. RYTHER : The annual cycle of primary production in the Sargasso Sea off Bermuda 351
Book Reviews : 368
Errata : 368

Author Index

(B) denotes Book Review

ABYZOV, S. S. p. 335.
ARONS, A. B. pp. 140, 217.
ARTHUR, R. S. p. 287.
BAKER, A. de C. p. 206.
BRATTSTRÖM, H. p. 75 (B).
CHARNEY, J. G. p. 303.
CORWIN, N. p. 60.
DANSGAARD, W. p. 346.
DEFANT, A. p. 262 (B).
FOLSOM, T. R. p. 169.
FØYN, E. p. 368.
GUILLARD, R. R. L. p. 65.
HAMON, B. V. p. 155.
HASLE, G. R. p. 38.
HAYTER, P. J. D. p. 184.
HERSEY, J. B. p. 170.
HILL, M. N. p. 193.
HUBBARD, C. J. p. 239.
ICHIYE, T. p. 16.
JOUSÉ, A. P. p. 187.
KETCHUM, B. H. p. 60.
KNAUSS, J. A. p. 265.
KRISS, A. E. pp. 88, 173, 335.
LACOMBE, H. p. 211.
LAUGHTON, A. S. pp. 259, 260 (B).
LEBEDEVA, M. N. p. 173.
LISITIZIN, A. P. p. 77.
MATTHEWS, D. H. p. 261 (B).
MCGILL, D. A. p. 60.
MCGOWAN, J. A. p. 125.
MEAD, G. W. p. 234.
MENZEL, D. W. p. 351.
MITZKEVICH, I. N. pp. 173, 335.
NEUMANN, G. p. 328.
NORDSTROM, S. G. p. 169.
RICHARDSON, W. S. p. 239.
ROBINSON, M. K. p. 249.
ROBINSON, A. R. p. 311.
RYTHER, J. H. pp. 65, 72, 351.
SOKOLOVA, M. N. p. 1.
STEELE, J. H. p. 69.
STOMMEL, H. pp. 140, 217, 298.
SWALLOW, J. C. p. 155.
THOMAS, C. W. p. 5.
VERONIS, G. p. 318.
WOLLF, T. p. 95.
WOOSTER, W. S. p. 263.
WÜST, G. p. 245.
YENTSCH, C. S. p. 72.
ZHIVAGO, A. V. p. 77.

Vol
6
1959

On the distribution of deep-water bottom animals in relation to their feeding habits and the character of sedimentation

M. N. SOKOLOVA

(*Received* 14 *April* 1958)

Abstract—The distribution of bottom invertebrates correlated with their feeding habits were determined from an analysis of the contents of the digestive tracts of 83 species of animals collected in depths ranging from 1000 to 10,000 m. Most deep-sea bottom invertebrates are deposit-feeders (about 55 per cent), a smaller number (about 25 per cent) are suspension-feeders and still fewer (about 20 per cent) are carnivores.

The composition and quantitative abundance of deep-sea communities are largely determined by the amount of organic deposit on the surface of the sea floor, and by the quantity of organic matter in suspension in the water layers just above the bottom. Consequently, a zonation of the various communities arises, each such zone extending over regions characterized by a distinct type of sedimentation. The dominant organisms in each area are, thus, those best adapted to it by their feeding habits.

The relationship between the communities of bottom invertebrates and recent geological processes described for the Kurile-Kamchatka Trench are to be expected in all other deep ocean depressions under similar conditions.

THE trophic factor* may have a direct influence on the distribution of the deep-water bottom fauna and on the change in its composition with increasing depth. Deep-water bottom invertebrates collected during investigations on the *Vityaz* (1949-1955) in the north-west Pacific Ocean and in the abyssal area of the Bering Sea provided for the first time sufficient material for a study of their feeding habits. The contents of the digestive tracts of 1923 specimens belonging to 83 different species were examined (SOKOLOVA, 1956 ; 1956a ; 1957 ; 1957a ; 1958). These deep-water forms may be grouped according to their feeding habits : deposit-feeders, which ingest the surface sediments, suspension-feeders and carnivores (HUNT, 1925). Most are deposit-feeders (55 per cent), a considerable number are suspension-feeders (25 per cent) and the remainder carnivores (20 per cent). There are no herbivorous deep-water bottom invertebrates apart from the exceptions noted by MORTENSEN (1938). These feeding habits are similar to those of the shallow-water invertebrates of the same or closely related groups. Feeding habits, it is believed, may influence the distribution of the deep-water animals and the composition of the communities with increasing depths.

Because of the preponderance of deposit-feeders and suspension-feeders, the amount of organic matter suspended in the water close to the bottom and on the surface of the bottom may largely determine the abundance and kinds of bottom invertebrates in any particular area. The amount of organic matter in suspension and the rate of sedimentation depends on the hydrographic conditions of the water just above the bottom. The water movement there in turn depends on the bottom

* The concept, 'trophic factor' includes a complex of conditions, such as feeding habits of the animal, abundance, distribution and availability of food, etc., all of which determine the life habits of an animal, its distribution in space and several other important vital functions.

topography. On elevations and slopes, horizontal currents frequently transport the sediments elsewhere and hence sedimentation does not occur in such areas. In depressions and on level surfaces, on the other hand, sediments tend to accumulate.

A comparison of the data on the rate of sedimentation in the Kurile-Kamchatka Trench (BEZRUKOV, 1955) and in the Bering Sea (LISITZIN, 1955) with those for the abundance of bottom animals and their feeding peculiarities show that these are definitely interrelated. Where the sedimentation rate is rapid, the bottom animals are abundant and belong to all three feeding types, but are chiefly deposit-feeders. Over bottom, where the concentration of suspended matter is high, but where sedimentation is not great, the suspension-feeders form a large percentage of the benthos. Finally, where the rate of sedimentation is low and there is little suspended matter in the bottom water, there are far more deposit-feeders, living and feeding within the surface sediments and only a few of the other two groups. Furthermore, the bottom population in such areas is sparse.

Under certain conditions, the relationships described above may give rise to a zonal differentiation of the deep-water communities. In the north-western Pacific and the western trench of the Bering Sea, there are three zones with a definite quantitative relationship to the feeding habits of the benthic animals found there. These are dependent on an alternation of the slope of the bottom with increasing depth

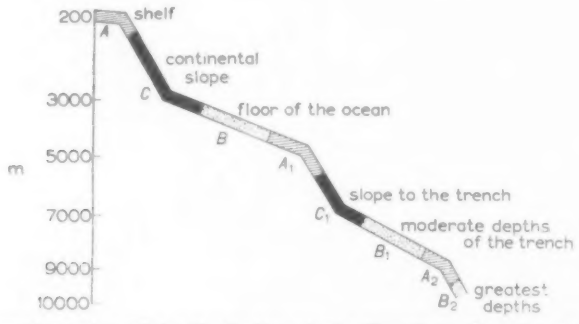

Fig. 1. Diagrammatic section of the Kurile-Kamchatka Trench to show the positions of the areas referred to in Table I and in the text.

(Fig. 1). Thus, at depths from 1000 to 10,000 m, there are two zones occurring at the base of the breaks in the slope with all three feeding types present, three zones on the breaks in the slope with suspension-feeders the most abundant and three on level bottom between one break in the slope and the next with deposit-feeders (Table 1). This last zone does not always occur and depends on a number of conditions, but chiefly on the sedimentation rate. A similar zonation of the bottom fauna also occurs in the Kurile-Kamchatka Trench " where there is a repeated alternation of linear zones of more or less active sedimentation and broken zones where there are no recent sediments " (BEZRUKOV, 1955, p. 111) and at other localities in the Bering Sea. Thus, the zonal distribution of the deep-water bottom fauna depends on zonal changes in sedimentation and the concentration of the suspended matter in the water just above the bottom.

As is noted by UDINTSEV (1955), alternating linear zones of active accumulation and erosion of the sediments are characteristic of all narrow trenches and island rises.

Vol 6 1959

In deep-sea trenches a close connexion exists between a large number of inter-related geological phenomena, including the topography, sedimentation and biological

Table 1. *Predominant organisms and their type of feeding found on the ocean bottom down to the greatest depths of the Kurile-Kamchatka Trench. The letters in the left-hand column refer to those shown on Fig.* 1

Area	*Predominant type of feeding*	*Predominant types of organisms*	*Ratio of weight of deposit feeders to that of suspension feeders*	*Ratio of weight of deposit feeders to that of carnivores*
A	suspension-feeders	Spongia	0·04	0·77
A_1		Coelenterata (Madreporaria Antipataria) Polychaeta Sabellidae	3·5	26·2
A_2		Pogonophora Grinoidea	2·5	27·5
B	deposit-feeders	Holothurioidea Elpidiidae Asteroidea Porcellanasteridae	45	116
B_1		Holothurioidea Elpidiidae Gephyrothuriidae	—	—
B_2		Holothurioidae Elpidiidae	84	162
C	deposit-feeders carnivores	Holothurioidea Molpadonia Crustacea Decapoda Coelenterata Actiniaria	17	1·0
C_1	suspension-feeders	Holothurioidea Molpadonia Asteroidea Porcellanasteridae Crustacea Isopoda Coelenterata Actiniaria	21	7·5

conditions determining the alternating faunal zones. Limited sampling of the bottom fauna from the Aleutian, Japanese, Mariana and Ryu-Kyu trenches indicate that the same zonation is characteristic of these areas.

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Vol
6
1959

Lithology and zoology of an Antarctic Ocean bottom core

CHARLES W. THOMAS

(*Received* 20 *October* 1958)

Abstract—Polar deep-sea cores provide material for a thermo-time scale.

An Antarctic Ocean deep-sea core from the Ross Sea can be correlated with three others from the same area, dated by Urry and described by HOUGH (1950). All the cores showed several alternations of glacial marine sediment and of fine-grained sediment which is apparently non-glacial. Organic deposits, notably radiolarian tests, are associated with glacial marine horizons and are lacking in the non-glacial ones, thus providing a zoological criterion for correlation. This association of sediments suggests inhibition of solar energy during deposition of strata where organic remains are lacking.

For the period of sedimentation represented by the core, a record of Pleistocene and recent thermal regimen is provided. Amelioration of climates over the past 6000 years is indicated in this record.

INTRODUCTION

THE core considered in this paper was taken on board the U.S. Coast Guard Icebreaker *Northwind* during the Byrd Antarctic Expedition of 1946-47. The author was Commanding Officer of this vessel. The work of core sampling was accomplished by Dr. JACK L. HOUGH of the Woods Hole Oceanographic Institution (now Professor of Geology, University of Illinois). He was assisted by Lieutenant WILLIAM G. METCALF U.S.N.R., also of the Woods Hole Institution.

Six cores were taken roughly along the 180th meridian and numbered N-3 to N-8 (inclusive). Of these cores N-3, N-4 and N-5 were analysed lithologically and described by HOUGH (1950). No work has been published on N-6 or N-7. The investigation of core N-8 was undertaken by the author. This paper concerns its analysis and correlation with N-3, N-4 and N-5.

Stations at which the foregoing cores were taken are as follows :

Core No.	*Length* (cm)	*Position*	*Depth* (m)
N-8	90	67°–00′S, 175°–00′W	3660
N-3	228	68°–26′S, 179°–12′W	3292
N-4	260	69°–12′S, 179°–45′E	3730
N-5	242	70°–17′S, 178°–00′W	2990

The position in which N-8 was taken, on January 31 1947, was free of sea-ice save for an occasional floeberg or glaçon. Icebergs were abundant. The remaining cores were taken in pack ice (HOUGH, 1950, Fig. 1).

Cores N-3 to N-5 were dated by Urry by the percentage of equilibrium method for uranium, ionium and radium.

PREVIOUS INVESTIGATIONS

PHILIPPI (1912) described the sediments collected by the *Gauss* of the German South Polar Expedition during the years 1901 to 1903, and used the term " glacial marine " to describe deposits of ice-rafted terrigenous sediments in the sea-bed.

Particle sizes of this material range from silt (4 μ in diameter) to gravel (greater than 2 mm in diameter).

While PIRIE (1913), NEAVERSON (1934) and STETSON and UPSON (1937) studied Antarctic sedimentation, no attempt was made to correlate past climatic phenomena with sedimentary horizons until HOUGH (1950). In the latters' description of Antarctic

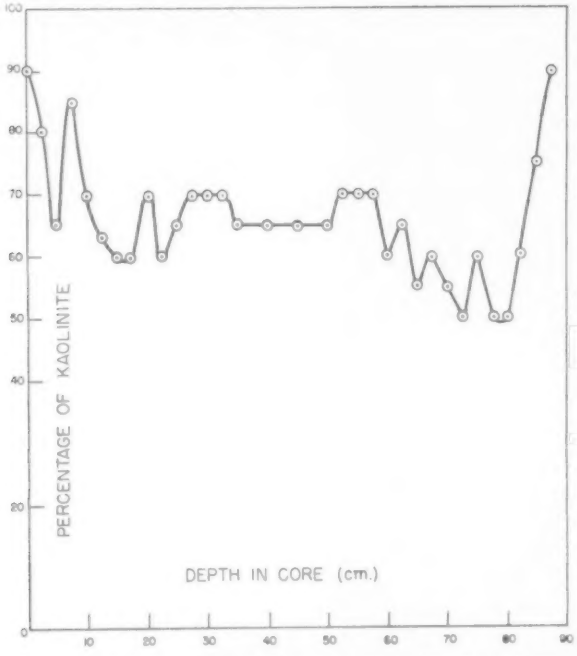

Fig. 1. Percentage of kaolinite in core N-8 relative to montmorillonite and illite as a function of depth in core.

cores (N-3, N-4 and N-5) he concluded that prior to 6000 years ago floating ice was apparently absent from the Ross Sea. This was based on the interpretation of an absence of glacial marine sediments as indicating a lack of icebergs by which terrigenous material is rafted. HOUGH believed these strata postulated warm conditions. LISITZIN (1958) emphasized the role of sea-ice as an agent of transport of terrigenous and eulittoral sediments in the Arctic and Antarctic Oceans.

METHODS OF STUDY

The *modus operandi* for collecting core material is explained by HOUGH (1950) in his description of cores N-3, N-4 and N-5.

The 90 cm length of core N-8 was divided into segments of 2·5 cm each. From each segment, half the material was wet-sieved through a 0·031 mm screen and thoroughly mixed, to produce a representative vertical section.

The relative percentage of major clay groups found in the core was determined by the acid-mineral stain method. This method was checked by a few pilot samples using the thermal-differential method. The results agreed.

Material from each segment was spread uniformly over a tray so as to constitute a single layer representing a vertical section of the zone and examined microscopically. These samples were not acid-treated in order to avoid destruction of calcareous tests, skeletal parts and clastics.

An accurate count of the principal constituents of the organic sediments (radiolaria and coccoliths) was taken and representative specimens of radiolaria set aside for identification. Radiolarians per cm^2 spread over the surface of the tray were counted and recorded.

The incidence of radiolaria and coccoliths appears in Table 2.

DISCUSSION

Clastic sediments considered in this discussion are divided into four general classes:

(1) Clay or fine-grained, well sorted particles smaller than silt (not glacial marine).

(2) Silty particles of fine-grained to coarse material (glacial marine).

(3) Glacial marine sediment composed of medium (250 μ) to coarse (500 μ) particles or grains.

(4) Pebbles, more than 500 μ: In core N-8, the largest weighed 650 mg.

In the glacial marine sediment, the grains were found to be unweathered, angular to subangular particles predominantly of quartz but including calcite, biotite, magnetite and feldspar as well as some metamorphosed and unclassified material. Pebbles were mainly unweathered, composed chiefly of quartz or metasedimentary complexes. Most of them were coated in whole or in part by a thin layer of black iron and manganese oxide. The silt-sized particles consisted principally of calcite and biotite.

Radiolarian tests (which are siliceous) were probably the most important organic constituents of the sediments. These were abundant at each end of the core (Fig. 3), lacking at the middle portion, and seemingly undifferentiated as to time of deposition. The same incidence was comparable for coccoliths and diatoms. Included in the faunal horizons but not specifically noted was a variety of siliceous sponge spicules, whose deposition seemed also concurrent with radiolarian tests. No foraminiferan shells were present in this core ; probably because the ocean depth at 3660 m is undersaturated with calcium.

Because of vertical mixing in zones of the Antarctic Convergence and Divergence (near the continental periphery) and lateral vectors of transport between them, the Antarctic Ocean is rich in silicates. This engenders an abundant population of diatoms (basic food supply of the ocean), and also explains the copious radiolarian population.

The problem herein required reduction of sedimentation from a geometric scale to a geologic one. N-3, for instance, covered a geometric depth of 228 cm and a geologic span of slightly over a million years. N-4, on the other hand, was 260 cm long and covered only about 187,000 years of time. In relating lithology to age in the foregoing cores, HOUGH (1950) remarked that the longest core (N-4) in 3730 m of water, contained material which was more rapidly deposited. This he attributed to the greater depth and its location in a closed depression in the bottom, where accumulation of sediments is more rapid.

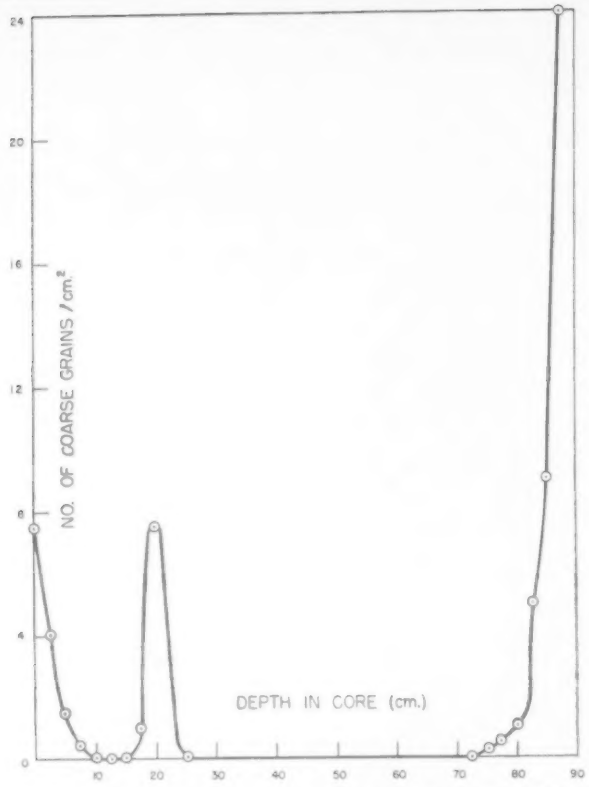

Fig. 2. Glacial marine sediments as a function of depth in core N-8.

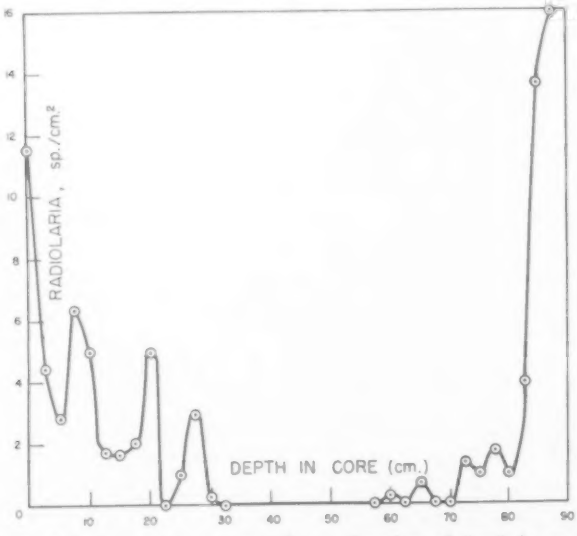

Fig. 3. Radiolaria specimens per cm^2 as a function of depth in core N-8.

A review of the bottom topography in the environment of the four cores shows N-3 to lie on the summit of a slight rise, with N-8 on the western slope of a swell. Hence, accumulation of sediments is not as rapid as at N-4. A plot was made of the number of grains per cm² of ice-rafted sediment (Fig. 2) as a function of depth in the core, a method similar to that used by EMILIANI (1955). This provided a basis by which lithologic horizons of N-8 were correlated with those of N-3, N-4 and N-5. Allowing for compaction, this correlation shows the 90 cm length of N-8 has a geologic equivalent of approximately 115,000 years (Fig. 4).

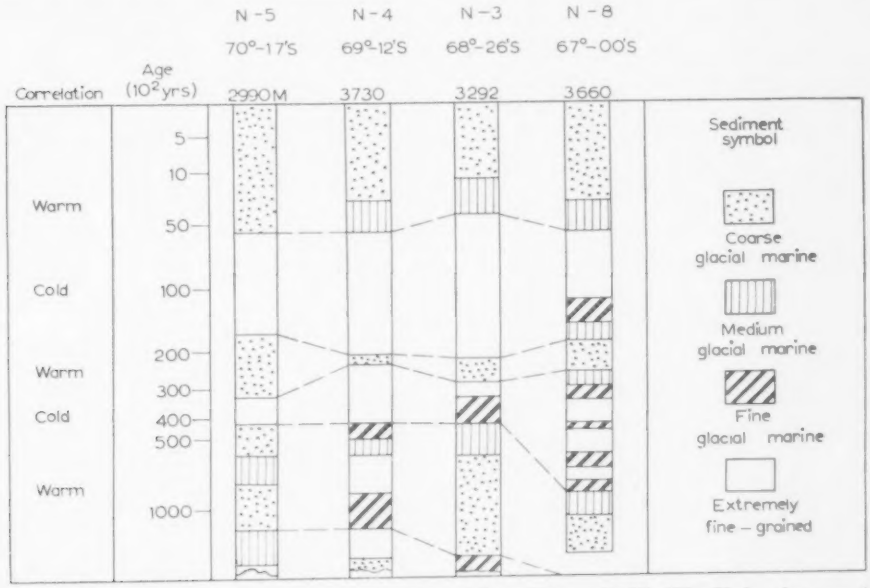

Fig. 4. Lithology of Ross Sea, Antarctic cores adapted from HOUGH (1950), [Fig. 2] showing correlation of core N-8 to dated cores, N-3 to N-5. Basal part of N-3 to N-5 not shown.

Table 1 shows the relative abundance of the three major clay groups. A plot of the ratio of kaolinite, the dominant member (Fig. 1) shows that during periods of sedimentation by ice transport, its percentage increased while that of the other members, montmorillonite and illite diminished*.

An examination of Tables 1 and 2 shows a uniform relationship between the incidence of radiolarian tests, coccoliths and ice-rafted clastics. Although no record was made of diatomaceous material, its occurrence was generally consonant with these sediments.

It is significant that organic horizons of N-8 correlate with its own lithology and that of N-3, N-4 and N-5, inasmuch as during periods of maximum sedimentation by ice-rafting, skeletal remains of organisms are most abundant. On the other hand, when glacial marine sediments are lacking there is an absence or paucity of organic sediments. This suggests that some mechanism must have inhibited or arrested organic production during periods when sedimentation by ice-rafting was denied.

*Similar results were obtained in the analysis of clay groups in IGY Core A-2 from the Arctic Basin.

One of the most elementary reasons for the foregoing phenomena which readily comes to mind is the possibility of an absence of food by which populations may be sustained. This would appear to be related to basic food production; fundamentally diatoms, also other algae and dinoflagellates.

There is no evidence of any material change in the circulation of atmosphere in the Antarctic or sub-Antarctic regions; hence, there could have been no significant shift in the circulation of the Ross Sea sector of the Antarctic Ocean. Therefore, a lack of organic production cannot be attributed to a paucity of nutrients. It appears most likely that solar energy was not available for photosynthesis.

Table 1. *Lithologic Table.*

Showing relationship of clastics and percentages of major clay groups to depth in core N-8

	Depth in core (cm) Percentages												
Major clay groups ..	0·00	2·50	5·00	7·50	10·00	12·5	15·0	17·5	20·0	22·5	25·0	27·5	30·0
Kaolinite	90	80	65	85	70	65	65	60	70	60	65	70	70
Montmorillonite ..	10	12	20	13	25	25	35	35	25	35	30	25	24
Illite	0	8	15	2	5	10	5	5	5	5	5	5	1
Pebbles (mg) ..	600	400							40				
Coarse grains/cm² ..	7·5	4·0	1·5	0·25				1·00	7·50	3·00			
Fine grains/cm² ..				1·25			0·10	1·00	9·00	3·00	0·30	6·00	0·50
Major clay groups ..	32·5	35·0	37·5	40·0	42·5	45·0	47·5	50·0	52·5	55·0	57·5	60·0	62·5
Kaolinite	70	65	65	65	65	65	65	65	70	70	70	60	65
Montmorillonite ..	25	30	25	25	20	25	25	20	25	30	20	20	15
Illite	5	5	10	10	15	10	15	15	5	0	10	20	20
Pebbles (mg).. ..													
Coarse grains/cm² ..													
Fine grains/cm² ..		7			4	0·4	2·0	1·0			0·5	2·0	1·0
Major clay groups ..	65·0	67·5	70·0	72·5	75·0	77·5	80·0	82·5	85·0	87·5	90·0		
Kaolinite	55	60	55	50	60	50	50	60	75		90		
Montmorillonite ..	25	25	25	30	20	25	30	25	23		10		
Illite	20	15	20	20	20	25	20	15	2		0		
Pebbles (mg) ..					40	80	70	160	20		650		
Coarse grains/cm² ..				0·05	0·25	0·50	1·0	5·0	9·0		25·0		
Fine grains/cm² ..	1·0	0·25	0·05	1·0	2·0	1·75	2·0	4·0	7·0		10·0		

Table 2. *Incidence of radiolaria species with depth in core N-8*

Specimen	*Depth in core (cm) specimens/cm²*								
	0·00	2·50	5·00	7·50	10·0	12·5	15·0	17·5	20·0
Cenosphaera favosa Haeckel ..	1·5	0·25	0·25	1·0	0·70			0·25	1·0
Cenosphaera cristata Haeckel ..	2·0	0·50	0·10	0·5	0·25			0·25	
Dorylonchidium sp.									
Stylatractus neptunus Haeckel ..			0·25		1·0				
Cenellipsis oblonga Haeckel ..									
Cenellipsis faceta Haeckel									
Spongodruppa terebintha Haeckel ..									
Sethodiscus phacoides Haeckel ..	1·5			2·0			0·5	0·25	2·0
Spongodiscus favus Ehrenberg var *maxima* Popofsky.. ..	2·0	1·30	1·2	2·3	2·0	1·0	1·0	0·40	2·0
Spongodiscus sp.	1·3	1·0					0·25	0·10	
Stylotrochus craticulatus H. ..	0·25	0·20						0·10	
Spongotrochus sp.	2·50	1·0	1·0	1·0	1·0	0·75		0·25	
Dictyocephalus sp. A.		0·2							
Dictyocephalus sp. B.	0·25								
Tripodiscum tristylospyris H. ..	0·1				0·25				
Dicotocapsa microcephala H. ..							0·4		
Coccoliths (approximate No.) ..	60	110	30	80	50	30	15	5	2

Vol 6 1959

Table 2 continued

Specimen	*Depth in core (cm) specimens/cm²*								
	22·5	25·0	27·5	30·0	32·5	35·0	37·5	40·0	42·5
Cenosphaera favosa Haeckel ..		1·0	0·5						
Cenosphaera cristata Haeckel ..									
Dorylonchidium sp.									
Stylatractus neptunus Haeckel ..									
Cenellipsis oblonga Haeckel ..									
Cenellipsis faceta Haeckel									
Spongodruppa terebintha Haeckel									
Sethodiscus phacoides Haeckel ..	0·10								
Spongodiscus favus Ehrenberg var. *maxima Propofsky* ..	0·50	0·25							
Spongodiscus sp.									
Stylotrochus craticulatus H. ..									
Spongotrochus sp.	0·10								
Dictyocephalus sp. A.									
Dictyocephalus sp. B.									
Tripodiscum tristylospyris H. ..									
Dicotocapsa microcephala Haeckel									
Coccoliths (approximate No.) ..	3								

	45·0	47·5	50·0	52·5	55·0	57·5	60·0	62·5	65·0
Cenosphaera favosa Haeckel ..							0·15		
Cenosphaera cristata Haeckel ..							0·10		
Dorylonchidium sp.									
Stylatractus neptunus Haeckel ..									
Cenellipsis oblonga Haeckel ..									
Cenellipsis faceta Haeckel									
Spongodruppa terebintha Haeckel									
Sethodiscus phacoides Haeckel ..									
Spongoidiscus favus Ehrenberg var. *maxima* Popofsky ..									
Spongodiscus sp.									
Stylotrochus craticulatus H. ..									
Spongotrochus sp.									
Dictyocephalus sp. A.									
Dictyocephalus sp. B.									
Tripodiscum tristylospyris H. ..									
Dicotocapsa microcephala H. ..									
Coccoliths (approximate No.) ..									

	67·5	70·0	72·5	75·0	77·5	80·0	82·5	85·0	87·5
Cenosphaera favosa Haeckel ..					0·20		1·0	2·1	3·2
Cenosphaera cristata Haeckel ..					0·10		1·5	1·4	3·0
Dorylonchidium sp.								2·5	
Stylatractus neptunus Haeckel ..									
Cenellipsis oblonga Haeckel ..									2·0
Cenellipsis faceta Haeckel									0·10
Spongodruppa terebintha Haeckel								1·2	
Sethodiscus phacoides Haeckel ..				0·25	0·25				4·0
Spongodiscus favus Ehrenberg var. *maxima* Popofsky ..								2·0	1·75
Spongodiscus sp.							0·50	1·5	0·25
Stylotrochus craticulatus H. ..									0·20
Spongotrochus sp.					1·35	0·75	1·0	3·0	1·0
Dictyocephalus sp. A.									0·10
Dictyocephalus sp. B.									
Tripodiscum tristylospyris H. ..									0·10
Dicotocapsa microcephala H. ..						0·10			0·10
Coccoliths (approximate No.) ..			3	1	0	18	30	20	10

In a rather intimate association with sea-ice over the past sixteen years and earlier, the author has often observed a condition which exists in the winter and spring in the higher latitudes which can best be described as 'frozen solid.' When temperatures are lowered materially below freezing, sea-ice integrates all floating material into a consolidated pack that is usually shorefast. Icebergs and floebergs are held immobile in the grip of a blanket of sea-ice. This condition exists seasonally in both polar regions, the amount of snow cover being materially greater in the Antarctic. If ice which is, or has been, in contact with sediments is denied mobility, vehicles for the transport of terrigenous or eulittoral material are unavailable.

During the Antarctic winter and spring the pack spreads north of the position of core N-8. It is not difficult to visualize that with a drop of only a few degrees in the mean annual temperature, the entire Antarctic Ocean* could be completely frozen the year round with a blanket of snow and sublimated material covering the surface. Such a cover could well interpose an insulating medium which would materially inhibit the transfer of solar radiation from atmosphere to sea-water. One is led to speculate then, as to whether or not sufficient energy is available beneath this medium to support photosynthesis by holophytic plankton – the basic food of the sea.

The average thickness of Antarctic sea-ice is 2 or 3 m with a sublime-snow cover of around 50 cm or more (author's observation). If this be so to-day, these media must have been equally thick under colder conditions.

To determine the amount of energy which penetrates ice and snow of the foregoing order of magnitude, it is necessary to make a theoretical computation based upon empirical data.

The amount of direct solar radiation which penetrates ice is a function of the coefficient of extinction of the medium. This coefficient is specific to each medium and may be plotted as a curve: $1/M \log I = S$.

$$\text{or} \quad \frac{dI}{dS} = \frac{I}{K} \tag{1}$$

$$\log I = KS \tag{2}$$

$$I\lambda = I_0 e^{K\lambda S}. \tag{3}$$

Where : I is intensity of energy

$I\lambda$ is ultimate intensity/wavelength

I_0 is residual intensity after reflexion at surface

$K\lambda$ is coefficient of extinction/wavelength

S is depth in medium.

I_0 available for transmission (in percentages) is dependeat upon the albedo. From basic data derived by LILJEQUIST (1956) the albedo in snow-covered regions of 67°S with computed solar altitude of 41° is an approximate average minimum of 78 per cent which is valid through the month of December. (It will be recalled that 67°S is the latitude of Core N-8). Moreover, the coefficients of extinction computed by LILJEQUIST for snow in 71°S are equally applicable to 67°S. The bands of the spectrum that

*That body of water which lies south of the Antarctic Convergence.

Vol 6 1959

must be considered are those which play a significant role in photosynthesis. The coefficients are for standard snow between November 20 and December 7 for various conditions of cloudiness normally encountered. They are only slightly different (within a small fraction) from a mid- to late December value (for which no empirical data are available).

Colour	Deep Blue	Light Blue	Orange	Red
Filter	BG-12	BG-9	OG-1	RG-2
	0·42 μ	0·52 μ	0·59 μ	0·65 μ
K (mean)	0·066	0·083	0·114	0·172

Using the foregoing values in equation (3) with an initial intensity of 22 per cent of incident radiation through 50 cm of snow cover, one obtains at 50 cm depth :

	Deep Blue	Light Blue	Orange	Red
$I\lambda$	0·0077	0·0037	0·00066	0·0000

LILJEQUIST did not derive extinction coefficients for various types of ice. Hence, to obtain coefficients for ice, values determined empirically by SAUBERER (1938) for glacial ice are used. By plotting, extrapolating and substituting in depth one derives values for K for appropriate bands of the spectrum. Sea-ice in the Antarctic is, on the whole, wind-blown. Coefficients of extinction for wind-blown ice are calculated as:

	Deep Blue	Light Blue	Orange	Red
K (mean)	0·018	0·021	0·029	0·043

Beneath 2·0 m of wind-blown ice and 50 cm of sublimated material and snow cover, a mean of 0·00018 of incident intensity should reach sea-water. With a maximum solar altitude of 41° (for 67°S latitude) approximately 1·15 g cal/cm²/min reach the earth. Through the foregoing media then, the amount of energy penetrating during a period of maximum solar radiation in 67°S would equal 0·00021 lys/min or about 1·2 ft candles. According to HOLMES (1957), SCHREIBER (1927) determined compensation (threshold) requirements in energy for photosynthesis by *Biddulphia mobilensis* (a diatom) and found it to be 10 ft candles. While requirements of only a few diatoms, dinoflagellates and algae are known, this apparently is the lowest on record.

CONCLUSION

From the foregoing analysis it appears that production of a basic holophytic food supply cannot take place in areas perpetually covered with sea-ice and snow – where a 'frozen solid' condition exists. This is because solar radiation cannot penetrate the solidified media in sufficient quantities per unit time.

Horizons of N-8 in which ice-rafted sediments occur are characterized by a representation of skeletal organic remains. The concurrent incidence of these sediments suggest an absence of a blanket of sea-ice at least part of the time, with freedom of bergs to move across the area and deposit terrigenous material.

Horizons lacking in ice-rafted and organic sediments can be attributed to cold climatic conditions. Had the seas represented by them been open, organic production would have proceeded with unimpeded prolificness, which the sediments would reflect.

During ' warm ' conditions the relative abundance of kaolinite was high in core N-8. During ' cold ' periods it decreased while that of montmorillonite increased. Since this relationship was noted only in the analysis of core N-8 plus one from the Arctic Basin, more data are needed before drawing conclusions regarding its significance.

In general, the following sequence of climatic conditions is inferred for the Ross Sea region through correlation of material in cores N-8, N-3, N-4 and N-5. The terms ' warm ' and ' cold ' do not necessarily imply optimal conditions :

Present to 6000 years ago	warm
6000 to 18,000 years ago	cold
18,000 to 32,000 years ago	warm
32,000 to 42,000 years ago	cold
42,000 years ago to earlier (around 100,000)	warm

The foregoing climatic inferences are generally consonant with conclusions of SAKS *et al.* (1954) who analysed an unstated number of cores taken by Soviet high latitude expeditions in the Arctic Basin*.

*The following sequence was postulated :

Present to 9-10 thousand years ago	warm
9-10 thousand to 18-20 thousand years ago	cold
18-20 thousand to 28-32 thousand years ago	warm
28-32 thousand to 45 thousand years ago	cold
45-50 thousand years ago	warm

The ' climatic optima ' which HOUGH (1950) believed to exist at times when glacial marine material was not deposited were really ' cold ' periods. At such times there may have been a retreat of terrestial glaciers whilst sea-ice blanketed the polar seas.

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Vol 6 1959

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Vol.
6
959-60

On long waves in a stratified, equatorial ocean caused by a travelling disturbance

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Abstract—Long waves, including Rossby waves and gravity waves, caused by a meteorological disturbance in an equatorial region are discussed by use of a theoretical model of a two-layer ocean. With the assumption that the frequency of wind stress variation is larger than the Coriolis' parameter, wave equations with a wind stress divergence as a forcing function are obtained. Forced waves initiated by a westward moving disturbance similar to the equatorial easterly waves indicate a resonance in the frequency range of gravity waves in a baroclinic mode and also in that of Rossby waves in both modes. The resonance in the gravity wave range seems to be responsible for the pronounced 4-day period oscillation detected at some Pacific islands.

Waves initiated by a sudden wind show no resonance, but interference between forced and free waves occurs near the period of resonance. The effect of a linear frictional force is shown to be more important in the Rossby wave range than in the gravity wave range. The effect of coasts is also studied and the proper oscillation of an ocean between two coasts is found to occur both in the Rossby wave range and gravity wave range.

When the frequency of the waves is decreased, the north-south distribution of the amplitudes must be modified owing to an increasing effect of the Coriolis' force in the sub-tropical zone of the wind. Analysis of the effect of the South Equatorial Current indicated that the response of sea-level has another resonant period in the intermediate range between Rossby waves and gravity waves. On the other hand, the current makes the resonant spectra of a period in Rossby waves so sharp that the resonance may be considered for practical purposes to vanish. The current also increases the amplitude of waves with periods of several days.

(1) INTRODUCTION

THE response of the ocean in an equatorial region to a changing wind system is particularly interesting theoretically for two reasons. First, in such a region the geostrophic approximation cannot be used, although it is the dominant relation in most large-scale atmospheric or oceanic waves in extra-tropical latitudes. The main terms in balancing a pressure gradient near the equator are the inertia terms and perhaps a frictional force rather than the Coriolis' force. Secondly, in the equatorial region over the ocean there is a meteorological disturbance moving westward, called easterly waves (PALMER, 1952). Such a wind system causes a resonance in Rossby waves (VERONIS and STOMMEL, 1956 ; ICHIYE, 1958b) in other latitudes.

How, then, are equations for long waves modified in the equatorial region without using the geostrophic assumption ? Is there resonance in a wave system in the equatorial region initiated by a westward moving disturbance ? The mean sea-level at Canton Island and other stations near the equator in the Pacific shows a remarkable 4-day period, seeming to correspond to the variation of the north-south component of the wind caused by easterly waves (GROVES, 1957).

The present work began primarily as a result of theoretical curiosity. The theoretical models are chosen to simulate the actual situation as closely as possible,

Vol
6
1959

in order to compare the results with observed facts, unless such models become too complicated mathematically. A two-layer ocean was considered, since, as is shown later, the density stratification influences the resonant condition in the equatorial region. In addition, the relation corresponding to the geostrophic approximation is derived with the assumption that the square of frequency of the waves is negligible compared with that of Coriolis' parameter. The perturbation equations for sea-level and inter-surface are obtained from the vorticity equation. These wave equations differ from those of other latitudes, where the geostrophic relation is valid, in the divergence term giving gravity waves rather than inertia waves, and also in term for the external force which mainly consists in divergence of wind stresses rather than their curl. These equations are valid only very near the equator for a relatively high frequency range, corresponding to gravity waves. In the sub-tropical side of the wind system, the geostrophic assumption is valid, so that the solution of the above equations must be modified. Such a modification is discussed below (Section 7).

The resultant wave equations are solved by separating them into two normal modes. The free waves represented by these equations consist of Rossby waves and gravity waves.

There is almost no interaction between these two wave ranges in either mode. There are two kinds of resonance from a westward moving disturbance : one in the Rossby wave range, the other in the gravity wave range. In a barotropic mode the latter period of resonance is very short and less than one day ; but for a baroclinic mode it is about 4 or 5 days, corresponding to the observed 4-day period of mean sea-level. This correspondence is further favoured when we take into account the effect of a frictional force or the existence of a mean current flowing westward, as the South Equatorial Current.

In examining the effect of coasts on the waves, it is found that there is another range of frequency of proper oscillation of the ocean between west and east coasts due to the reflexion of the gravity waves. Such a proper oscillation also causes marked fluctuations in the amplitude of free waves initiated by a disturbance in a bounded ocean. This paradoxical result can be avoided by introducing a term for the frictional force as in the case of longer Rossby waves in a bounded ocean (ICHIYE, 1958b). Sometimes a moving wind system appears suddenly. Analysis of waves caused by such a transitional system indicates that at the resonant period forced waves are cancelled by free waves with that period ; but that there is an interference between these two waves having relatively large amplitudes near the resonant period.

(2) DERIVATION OF WAVE EQUATIONS IN THE EQUATORIAL REGION

As in a previous paper (ICHIYE, 1958b), a two-layer ocean is considered and the equations of motion and continuity in both layers are given by

$$\mathbf{P}\, u - fv = -g\partial\,(\eta + \eta')/\partial x + \tau_x/h \tag{1}$$

$$\mathbf{P}\, v + fu = -g\partial\,(\eta + \eta')/\partial y + \tau_y/h \tag{2}$$

$$\partial\eta/\partial t + h^{-1}\,(\partial u/\partial x + \partial v/\partial y) = 0 \tag{3}$$

$$\mathbf{P}'\, u' - f v' = -g\partial\,(\xi\eta + \eta')/\partial x \tag{4}$$

$$\mathbf{P}' v' + fu' = -g\partial(\xi\eta + \eta')/\partial y \tag{5}$$

$$\partial\eta'/\partial t + (h')^{-1}(\partial u'/\partial x + \partial v'/\partial y) = 0 \tag{6}$$

where x and y are horizontal co-ordinates for east and north, respectively; u, v and τ_x, τ_y are velocity and wind stress components of x and y direction ; h is the depth of the upper layer ; k is a friction coefficient ; and ξ is the ratio of density ρ/ρ', with the prime indicating corresponding quantities in the lower layer.
Further $\mathbf{P}$ and $\mathbf{P}'$ mean the operator

$$\mathbf{P} = (\partial/\partial t) + k, \quad \mathbf{P}' = (\partial/\partial t) + k' \tag{7}$$

From (1) and (2) we have

$$(\mathbf{P}^2 + f^2)\, u = \mathbf{P}\, H_x + fH_y \tag{8}$$

$$(\mathbf{P}^2 + f^2)\, v = \mathbf{P}\, H_y - fH_x \tag{9}$$

where

$$H_x = -g\,\partial(\eta + \eta')/\partial x + \tau_x h^{-1}, \quad H_y = -g\partial(\eta + \eta')/\partial y + \tau_y h^{-1} \tag{10}$$

Similar equations are derived from (4) and (5) with all quantities with a prime, and instead of (10) H_x' and H_y' are given by

$$H_x' = -g\partial(\xi\eta + \eta')/\partial x, \quad H_y' = -g\partial(\xi\eta + \eta')/\partial y \tag{11}$$

When we consider a periodic motion and for the time neglect frictional force, we can put $|\mathbf{P}| = \nu$ where ν is a frequency of the motion. Then the ratio $f|\mathbf{P}|^{-1}$ can be expressed by more intuitive quantities such as a latitude ψ of the considered region and the period n of the motion in days, as $f\nu^{-1} = n \sin\psi$. In other words, at a latitude ϕ, ν exceeds f for the motion with a period shorter than cosec ψ days, and such critical periods are as follows :

Latitudes (°)	20	10	8	5
Critical period (days)	2·92	3·86	7·19	11·47

At first we are concerned with the condition very near to the equator, so that it is justifiable to neglect f^2 against $|\mathbf{P}|^2$ in the left hand side of (8) and (9). Then we get the vorticity ζ from these two equations such as

$$|\mathbf{P}|^2 \zeta = h^{-1}\mathbf{P}\,\mathrm{curl}\,\tau + f\{g\nabla^2(\eta + \eta') - h^{-1}\,\mathrm{div}\,\tau\} - \beta H_y \tag{12}$$

$$|\mathbf{P}'|^2 \zeta' = fg\nabla^2(\xi\eta + \eta') - \beta H_y' \tag{13}$$

The vorticity equations in both layers are obtained by cross-differentiation of (1), (2) and (4), (5) and we have

$$\mathbf{P}\zeta - f\,\partial\eta/\partial t + \beta v = h^{-1}\,\mathrm{curl}\,\tau \tag{14}$$

$$\mathbf{P}'\zeta' - f\,\partial\eta'/\partial t + \beta v' = 0 \tag{15}$$

We obtain the perturbation equations by substituting (12) and (13) with (9) (neglecting f^2 in the left hand side) into (14) and (15) such as

$$\mathbf{P}(\nabla^2\overline{\eta + \eta'} - c^{-2}\mathbf{P}\,\partial\eta/\partial t) + \beta\,\partial(\eta + \eta')/\partial x = c^{-2}(\mathbf{P}\,\mathrm{div}\,\tau + \beta\tau_x) \tag{16}$$

$$\mathbf{P}'(\nabla^2\overline{\xi\eta + \eta'} - (c')^{-2}\mathbf{P}'\,\partial\eta'/\partial t) + \beta\,\partial(\xi\eta + \eta')/\partial x = 0 \tag{17}$$

Vol 6 1959

where

$$c^2 = gh, \quad (c')^2 = gh'.$$

In order to compare these equations with the quasi-geostrophic approximation which is valid in the extra-tropical region or for a longer period, we derive the similar perturbation equations to (16) and (17) by assuming $f^2 > \mathbf{P}^2$ in (18) and (19). Then the vorticity ζ is expressed by

$$\zeta = f^{-1}(g\,\nabla^2\overline{\eta + \eta'} - h^{-1}\,\mathrm{div}\,\tau + h^{-1}\,\mathbf{P}\,\mathrm{curl}\,\tau) - \beta f^{-2}(2H_x + H_y) \tag{18}$$

with similar relation for ζ' and the result is

$$\mathbf{P}\,\nabla^2(\eta + \eta') - \lambda^{-2}\,\partial\eta/\partial t + \beta\,\partial(\eta + \eta')/\partial x = c^{-2}(f\,\mathrm{curl}\,\tau + \mathbf{P}\,\mathrm{div}\,\tau + \beta\,\tau_x) \tag{19}$$

$$\mathbf{P}\;\nabla^2(\xi\eta + \eta') - (\lambda')^{-2}\,\partial\eta'/\partial t + \beta\,\partial(\xi\eta + \eta')/\partial x = 0 \tag{20}$$

where

$$\lambda^2 = c^2 f^{-2}, \quad (\lambda')^2 = (c')^2 f^{-2}.$$

By comparing these equations with (16) and (17) we can see at once that the divergence term $\partial\eta/\partial t$ or $\partial\eta'/\partial t$ produces gravity waves in the former system but gives inertia-gravity waves in the latter. Further, as for an external force causing wave motion, the divergence of the wind stresses is predominant in the former, while in the latter the curl of the wind stresses is dominant.

In deriving the vorticity from the equations (8) and (9) without any approximation on the left hand side, perturbation equations are obtained covering wider ranges of latitudes and frequency than those of the systems (16), (17) and (19), (20). However, these equations are too complicated for a clear insight into the physical properties of the waves. Therefore we consider here that the above two systems are valid, one near the equator and the other in the sub-tropical zone. The solutions for each system of equations in each region can be combined to describe the entire equatorial disturbance due to easterly waves which extend to the sub-tropical zone.

Since the system of (19) and (20) derived by the geostrophic approximation has been discussed elsewhere (CHARNEY, 1955 ; ICHIYE, 1958b), we will concentrate on the system of (16) and (17) which is valid only near the equator or for a frequency larger than that of ordinary Rossby waves.

(3) FREE WAVES IN A FRICTIONLESS OCEAN

For the time, let us neglect friction and the operator $\mathbf{P} \to \partial/\partial t$, so that the equations (16) and (17) can be solved by separation into normal modes (CHARNEY, 1955). Then we have

$$\frac{\partial}{\partial t}\left(\nabla^2 R_j - c_j^{-2}\frac{\partial^2 R_j}{\partial t^2}\right) + \beta\frac{\partial R_j}{\partial x} = c^{-2}\sigma_j\left(\frac{\partial}{\partial t}\,\mathrm{div}\,\tau + \beta\,\tau_x\right) \tag{21}$$

where the suffix $j = r$ and l means a barotropic and baroclinic mode, respectively. We put

$$c_j^2 = c^2\alpha_j^{-2}, \quad r = h/h', \quad \sigma_r \approx r, \quad \sigma_l \approx -1$$

$$\alpha_r^2 \approx r(1 + r)^{-1}, \quad \alpha_l^2 \approx (1 + r)(1 - \xi)^{-1} \tag{22}$$

We assume that R_j is proportional to $\cos(\omega t + \alpha x) \sin ly$ with $y = 0$ being at the equator. Then the frequency ω of free waves is a root of the following cubic equation.

$$\omega^3 - c_j^2 (\alpha^2 + l^2)\, \omega + c_j^2\, \alpha\, \beta = 0 \tag{23}$$

When we take physical and geometrical constants such as

$$h = 2 \times 10^4 \text{ (cm)}, \quad h' = 4 \times 10^5 \text{ (cm)}, \quad l = 2\pi \cdot 10^{-8} \text{ (cm}^{-1}\text{)}$$

$$\beta = 2{\cdot}3 \times 10^{-13} \text{ (cm}^{-1} \text{ sec}^{-1}\text{)}, \quad 1 - \xi = 2 \times 10^{-3}$$

in which the value of l is determined corresponding to a wind system of 2000 km in width, the first approximation of the roots of (23) becomes :

$$\omega_{1j} = \alpha\, \beta\, (\alpha^2 + l^2)^{-1}, \quad \omega_{2,3j} = \pm\, c_j\, (\alpha^2 + l^2)^{\frac{1}{2}}; \quad j = r, l \tag{24}$$

in which ω_{1j} is the same for a barotropic and baroclinic mode ; this corresponds to the period of Rossby waves. The roots $\omega_{2,3j}$ correspond to the gravity waves. The second approximation of these roots are given by

$$\omega_{1j} = \omega_1' + (\omega_1')^3\, (\omega'_{2j})^{-2}$$

$$\omega_{2,3j} = \omega'_{2,3j} - \tfrac{1}{2}\, \omega'_1. \tag{24'}$$

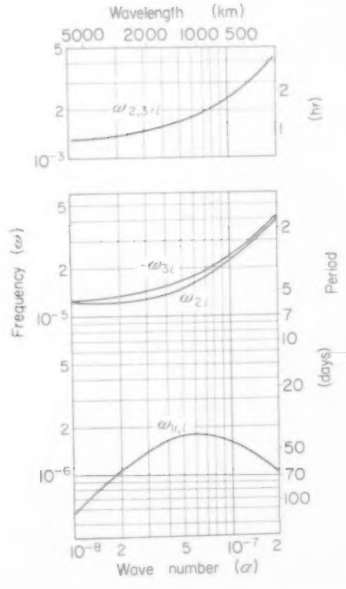

Fig. 1. Frequencies of free waves as functions of wave numbers.

where ω_1' and $\omega'_{2,3j}$ means the first approximation given by (24). For a range of α from 10^{-8} to 2×10^{-7} the first approximation (24) is sufficiently accurate for practical purposes except for the gravity waves of a baroclinic mode. This means that there is almost no interaction between Rossby waves and gravity waves with wave number α.

Vol 6 1959

In Fig. 1 these frequencies of free waves are shown as functions of a wave number α. In a barotropic mode the frequency $\omega_{2,3}$ is higher for gravity waves than for Rossby waves, so that the separation of these two waves is more complete than in a baroclinic mode. As expected from the equations (24), gravity waves are almost non-dispersive as a wave number α is higher than about 5×10^{-8}, while the Rossby waves become non-dispersive for a much smaller wave number.

(4) FORCED MOTION CAUSED BY A TRAVELLING WIND SYSTEM

In this section, the forced oscillation due to a wind system moving toward the west and corresponding to easterly waves in the equatorial region is considered. As pointed out by GROVES (1956), the variation of τ_x is very small compared with that of τ_y in the equatorial region, so that the divergence of wind stresses only need be considered. This also agrees with the model of easterly waves described by PALMER (1952). Thus we put

$$\operatorname{div} \tau = D \cos(\alpha x + \nu t) \sin ly \tag{25}$$

Then we get from the equation (20)

$$R_j = R_{0j} \cos(\alpha x + \nu t) \sin ly$$

$$R_{0j} = D\, \sigma_j\, c^{-2} (\nu^2 c_j^{-2} - \alpha^2 - l^2 + \alpha \beta \nu^{-1})^{-1} \tag{26}$$

The factor R_{0j} is computed as a function of frequency ν for a wave number $\alpha = 10^{-8}$ and 5×10^{-8}. The first wave number corresponds to a wavelength of 6283 km and the second to that of 1257 km. The observed wavelength of easterly waves is actually between 1500 km and 2000 km.

The surface elevation $\eta + \eta'$ and the deviation of the intersurface η' are computed

$$\eta + \eta' = (R_r - R_l)(1 + r)^{-1}, \quad \eta' = R_r (1 + r)^{-1} + R_l (1 - \xi)^{-1}$$

and also v_g and v'_g are defined by

$$v_g = \int g \{\partial (\eta + \eta')/\partial x\}\, dt = igl\, \nu^{-1} (\eta + \eta')$$

$$v'_g = \int g \{\partial (\xi\eta + \eta')/\partial x\}\, dt = igl\, \nu^{-1} (\xi\eta + \eta')$$

These are also considered to represent characteristic velocities in the upper and lower layer.

The numerical results are shown for values of ν from 10^{-6} to 10^{-4} in Figs. 2 and 3. From the equation (26), the amplitude becomes infinite at a frequency corresponding to that of free waves given as roots of (23). This is considered as a resonance, and particularly in the Rossby wave range such a resonance occurs only for a westward moving wind system (VERONIS and STOMMEL, 1956 ; ICHIYE, 1958b). In the present model this resonance occurs only for a wave number α of 5×10^{-8} in a considered range of frequency. On the other hand the resonance of gravity waves occurs both for a westward and for an eastward moving disturbance, since the frequency of free waves has both signs as shown in (24). However, in the model considered here, the resonant frequency is of an order of 10^{-3} in a barotropic mode. For a baroclinic mode, this resonant frequency is equal to $1{\cdot}48 \times 10^{-5}$ or the

resonant period is about 4·9 days for a wave number $\alpha = 5 \times 10^{-8}$. This seems to correspond to the predominant period of 4-days of mean sea-level fluctuation at Canton Island and other stations near the equator in the Pacific. (GROVES, 1957).

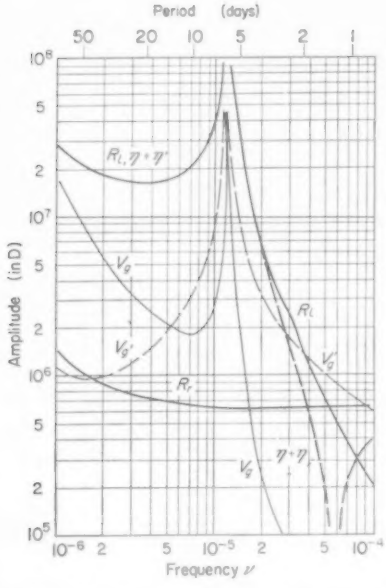

Fig. 2. Amplitudes of forced waves for $\alpha = 10^{-8}$ (heavy solid lines: R_r and R_l, heavy broken lines: $\eta + \eta'$, fine solid lines : V_g, fine broken lines : V_g'. V_g in m/sec, V_g' in cm/sec).

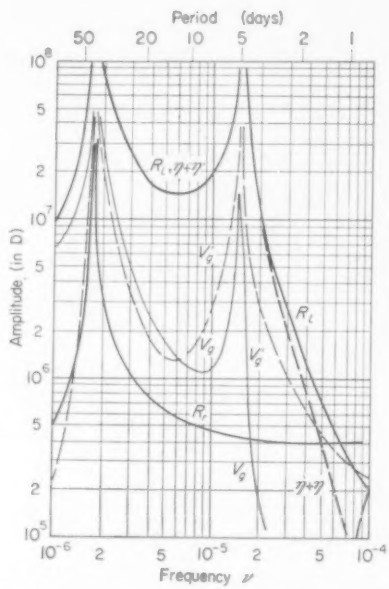

Fig. 3. Amplitudes of forced waves for $\alpha = 5 \times 10^{-8}$ (see legend to Fig. 2).

The resonant frequency in Rossby wave range is equal to $1{\cdot}78 \times 10^{-6}$ for $\alpha = 5 \times 10^{-8}$ and this range of ν is smaller than necessary for the assumption $\nu^2 > f^2$ so that it is doubtful whether this resonant frequency is exact, except very close to the equator.

(5) WAVES CAUSED BY A TRANSIENT WIND SYSTEM

When a wind system begins from a state of no motion, both free and forced waves occur simultaneously. After they are initiated, the free waves move with their proper speed (Section 3) and the forced waves proceed with the disturbance. Here we derive the relationships between free and forced waves caused by a sudden appearance of a moving wave system as expressed by the next relation

$$\text{div}\,\tau = 0, \quad t < 0\,; \quad \text{div}\,\tau = D\cos(\nu t + \alpha x)\sin ly, \quad t > 0 \qquad (27)$$

From the initial condition that at $t = 0$

$\eta + \eta' = 0$, $\eta' = 0$, and also

$\partial\eta/\partial t = \partial\eta'/\partial t = 0$

Vol 6 1959

which is derived from the equations of continuity (3) and (6), the initial condition for R_j becomes at $t = 0$

$$R_j = 0, \quad \partial R_j/\partial t = 0 \tag{28}$$

On the other hand, the fundamental equation (21) has a term $\partial\,(\mathrm{div}\,\tau)/\partial t$ in the right hand side and this becomes infinite at $t = 0$ for the condition (27). Therefore in the initial stage we should use the integrated form of (21) with time from $t = 0$ to $t = \epsilon$ such as

$$\nabla^2 R_j - c_j^{-2}\,\partial^2 R_j/\partial t^2 + \beta \int_0^{\epsilon} (\partial R_j/\partial x)\,dt = c^{-2}\,\mathrm{div}\,\tau.$$

Further, $\partial R_j/\partial x$ should be finite near $t = 0$, so that the integral $\int_0^{\epsilon} (\partial R_j/\partial x)\,dt$ should vanish as $\epsilon \to 0$. As a result, we have at $t = 0$

$$\nabla^2 R_j - c_j^{-2}\,\partial^2 R_j/\partial t^2 = c^{-2}\,\mathrm{div}\,\tau \tag{29}$$

and this relation should be satisfied as the third initial condition.

Then the general solution Rj of the equation (21) is given by

$$R_j = \mathbf{Re}\,(Ae^{i\omega_1 t} + Be^{i\omega_2 t} + Ce^{i\omega_3 t} + R_{j0}\,e^{ivt})\,e^{ilx}\sin ly \tag{30}$$

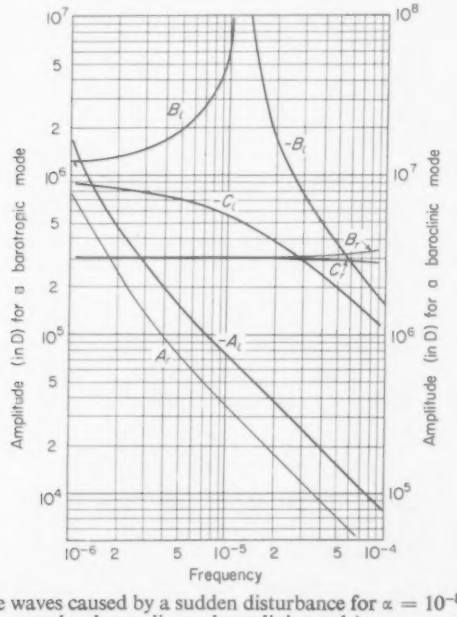

Fig. 4. Amplitudes of free waves caused by a sudden disturbance for $\alpha = 10^{-8}$ (fine lines : barotropic mode ; heavy lines : baroclinic mode).

where **Re** signifies the real part and A, B and C are complex constants to be determined from the initial conditions. Further ω_1, ω_2 and ω_3 are the frequencies of free waves given by the roots of (23) and R_{j0} is an amplitude of a forced wave given by (26). Since the solution (30) satisfies the equation (21) for $t > 0$ and it should be continuous with $t = 0$, the initial condition (29) becomes

$$(A/\omega_1) + (B/\omega_2) + (C/\omega_3) + (R_{j0}/\nu) = 0 \tag{31}$$

The constants A, B and C are determined from the initial conditions (28) and (31) and they are

$$A = \omega_1 (\nu - \omega_2)(\nu - \omega_3)\{\nu(\omega_1 - \omega_2)(\omega_3 - \omega_1)\}^{-1} R_{j0}, \text{ etc.} \tag{32}$$

$$A = D\sigma_j \alpha_j^{-2} \omega_1 \{(\nu - \omega_1)(\omega_1 - \omega_2)(\omega_3 - \omega_1)\}^{-1}, \text{ etc.} \tag{33}$$

Here we take ω_3 as negative, so that $\omega_3 = -\omega_2$. When $\nu = \omega_1$, the quantity R_{j0} becomes infinite (Section 4) but at the same time $A = -R_{j0}$ as seen from (32). The resultant solution (30) is finite, and this is true for the resonant period $\nu = \omega_2$.

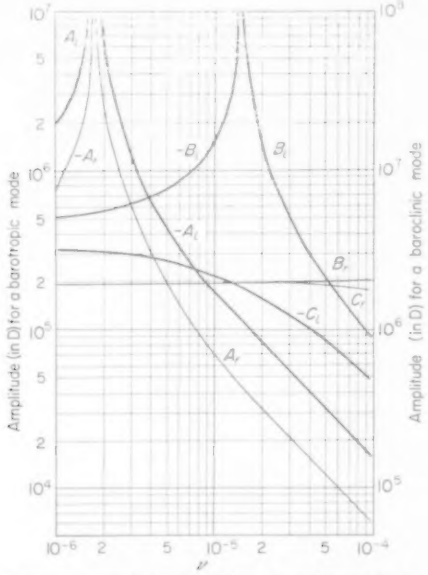

Fig. 5. Amplitudes of free waves caused by a sudden disturbance for $\alpha = 5 \times 10^{-8}$ (see legend to Fig. 4).

The values of A, B and C are shown as functions of ν for $\alpha = 10^{-8}$ and 5×10^{-8} in Fig. 4 and 5. From these figures it is seen that in a barotropic mode amplitudes of gravity waves B and C are almost constant for a range of ν between 10^{-6} and 10^{-5}, but for that of Rossby waves A decreases rapidly with a frequency higher than ω_1. On the other hand, in a baroclinic mode the amplitude B shows a resonance at ω_2 as expected from (33), but both B and C decrease more rapidly as frequency increases than in a barotropic mode. Generally, the values of A, B and C for a baroclinic mode are about ten times or more of the corresponding ones for a baroclinic mode except near resonant frequencies.

It is interesting to examine the solution near the resonant frequency ω_2, since this is very close to the observed dominant period of mean sea-level. When we put $\nu = \omega_2 + \epsilon$ with ϵ being very small compared with ν and $B = -(R_{j0} + \delta)$, then we have

$$B\cos(\omega_2 t + \alpha x) + R_{j0}\cos(\nu t + \alpha x) \approx -R_{j0}\sin(\omega_2 t + \alpha x)\sin \epsilon t$$

for a combination of free ω_2 — waves and forced waves. Therefore there are wave trains of a large amplitude B_{j0} which varies very slowly with a frequency ϵ, owing

Vol 6 1959

to the interference between the forced and free ω_2 — waves. This explains how a sudden appearance of easterly waves causes a large fluctuation with a period of 4 to 5 days. Of course, the present model is an ideal one, and there is still the possibility that a large fluctuation in amplitude is produced by the sudden formation of easterly waves.

(6) EFFECTS OF A FRICTIONAL FORCE AND COASTS

Knowledge of a frictional force acting on the wave motion is very incomplete and also an adequate boundary condition at a coast seems to be of non-linear character. Therefore the discussion in this section is limited to a linear friction only as already introduced in the fundamental equations in Section 3. Such a friction explains the basic mechanism for damping of wildly oscillating Rossby waves in a bounded ocean (ICHIYE, 1958b) or for interpretation of the observed damping of tsunami waves in shallow water (ICHIYE, 1958a), although it is mathematically simple.

In order to solve (19) and (20) in normal modes, we assume that $k = k'$. Then the equation of frequency of free waves becomes

$$(\omega - ki)\left\{c_j^{-2}\,\omega\,(\omega - ki) - l^2 - \alpha^2\right\} + \alpha\beta = 0 \tag{34}$$

instead of (23). The roots $\omega^{(2)}{}_{i,f}$ of (34) taking into account of the second approximation is given by

$$\begin{aligned}\omega^{(2)}{}_{1,f} &= (1 + \omega_1{}^3\ c_j^{-2}\,\alpha^{-1}\,\beta^{-1})\,\omega^{(1)}{}_{1,f}\\ \omega^{(2)}{}_{2,3,f} &= \tfrac{1}{2}\,ki \pm \left\{\tfrac{1}{4}\,k^2 + \omega_2{}^2 - \alpha\beta\,c_j{}^2\,(\omega^{(1)}{}_{2,f} - ki)^{-1}\right\}^{\frac{1}{2}}\end{aligned} \tag{35}$$

where the first approximation is

$$\omega^{(1)}{}_{1,f} = \omega_1 + ki, \quad \omega^{(1)}{}_{2,3,f} = \tfrac{1}{2}\,ki \pm (\omega_2{}^2 + \tfrac{1}{4}\,k^2)^{\frac{1}{2}} \tag{36}$$

with values of ω_1 and ω_2 given by (24).

However, correction terms in (35) are very small, because

$$\omega_1{}^3\,c_j^{-2}\,(\alpha\beta)^{-1} \leqslant 1{\cdot}1 \times 10^{-6} \quad \text{for} \quad j = r, \quad \leqslant 1{\cdot}3 \times 10^{-2} \quad \text{for} \quad j = l$$

$$\frac{\alpha\beta\,c_j{}^2\,\omega_2{}^{-2}}{|\omega_2{}^{(1)} - ki|} \leqslant 4{\cdot}3 \times 10^{-6} \quad \text{for} \quad j = r, \quad \leqslant 2{\cdot}6 \times 10^{-2} \quad \text{for} \quad j = l$$

for a range of α from 2×10^{-7} to 10^{-8}. Therefore, the first approximation (35) is sufficiently accurate. It is interesting that the decrease of free waves with time due to a friction is expressed as $\exp(-kt)$ for Rossby waves and $\exp(-kt/2)$ for gravity waves, even though the frequency of the former is far lower than that of the latter.

The forced oscillation with a friction is also expressed by

$$\begin{aligned}R_j &= \mathbf{Re}\left\{R_{0j}\exp i\,(\alpha x + \nu t)\right\}\sin ly\\ R_{0j} &= D\sigma_j\,c^{-2}\left\{c_j^{-2}\,(\nu - ki)^2 - (\alpha^2 + l^2)\,(\nu - ki)\,\nu^{-1} + \alpha\beta\nu^{-1}\right\}^{-1}\end{aligned} \tag{37}$$

instead of (26). The absolute value of R_{0j} is shown as a function of ν for $k = 10^{-7}$ and 10^{-6} with a wave number $\alpha = 10^{-8}$ and 5×10^{-8} in Fig. 6. As stated before, the effect of friction is more conspicuous in a smaller frequency range, and the resonance amplitude near ω_1 is much reduced by friction. On the other hand, in a gravity wave range its effect is quite slight, so that even for $k = 10^{-7}$ the resonance amplitude

near ω_2 is not much affected by friction. This condition also seems to be favourable in initiating a 4-day period resonance.

A more complete discussion of the effects of coastal boundaries on the Rossby waves has been presented elsewhere (ICHIYE, 1958b). Hence it is only briefly mentioned here. When the coasts exists, the solution of (21) including a linear friction is expressed by

$$R_j = \mathbf{Re}\,(A_1\, e^{i\delta_1 x} + A_2\, e^{i\delta_2 x} + R_{0j}\, e^{i\alpha x})\, e^{i\nu t} \sin ly \tag{38}$$

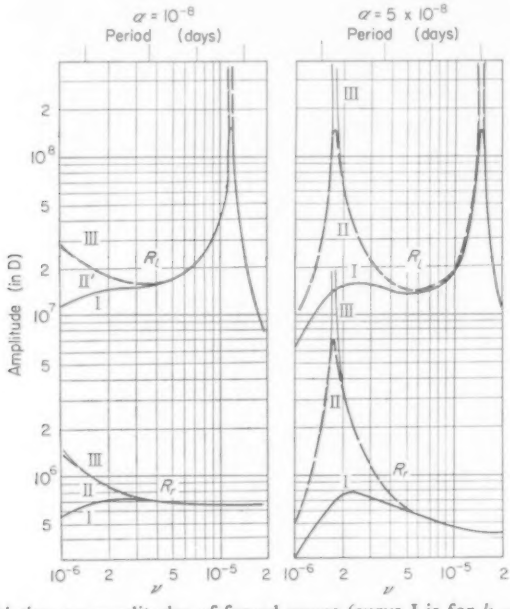

Fig. 6. Effect of a friction on amplitudes of forced waves (curve I is for $k = 10^{-6}$; curve II is for $k = 10^{-7}$; curve III is for $k = 0$).

for a divergence of wind stresses given by (25). In this equation R_{0j} is a term of the forced motion given by the equation (37) and δ_1 and δ_2 is the roots of the next quadratic equation.

$$\delta^2 - \beta\,(\nu - ki)^{-1}\,\delta + l^2 - c_j^{-2}\,\nu\,(\nu - ki) = 0 \tag{39}$$

The coefficients A_1 and A_2 are determined from the boundary conditions that

$$u = 0 \quad \text{at} \quad x = 0 \quad \text{and} \quad x = a.$$

When we assume that $\tau_x = 0$, as previously stated, then these conditions become

$$\partial R_j/\partial x = 0 \tag{40}$$

at $x = 0$ and $x = a$.

Putting

$$\delta_1 = p + q, \quad \delta_2 = p - q$$

with

$$p = \tfrac{1}{2}\,\beta\,(\nu - ik)^{-1} = p_1 + ip_2$$

$$q = \{p^2 - l^2 + c_j^{-2}\,\nu\,(\nu - ik)\}^{\frac{1}{2}} = q_1 + iq_2 \tag{41}$$

then we obtain from the conditions (40)

$$A_{1,2} = \tfrac{1}{2}\, i\, \alpha\, (e^{\mp iaq} - e^{i\alpha\overline{x-p}})\, (q \pm p)^{-1} \sin^{-1} aq\, R_{j0} \tag{42}$$

Without friction, p is a real number. On the other hand q is real only when $\nu > \nu_j$ or $\nu < \nu_{2j}$; otherwise it is imaginary, in which ν_{1j} and ν_{2j} are expressed by

$$\nu_{1,2j} = 2^{-\frac{1}{2}}\, c_j\, l\, \{1 \pm (1 - \beta^2\, c_j^{-2}\, l^{-4})^{\frac{1}{2}}\}^{\frac{1}{2}} \tag{43}$$

with ν_{1j} having the plus sign. Particularly for a barotropic mode, $\beta^2\, c_j^{-2}\, l^{-4} \ll 1$ so that the next approximation

$$\nu_{1j} = c_j\, l, \quad \nu_{2j} = \beta\, (2l)^{-1} \tag{43'}$$

is accurate enough. By taking the same values of constants as given in Section 3, we have

$$\nu_{1r} = 1{\cdot}32 \times 10^{-3}, \quad \nu_{2r} = 1{\cdot}83 \times 10^{-6}\ ; \quad \nu_{1l} = 1{\cdot}21 \times 10^{-5}, \quad \nu_{2l} = 1{\cdot}84 \times 10^{-6}.$$

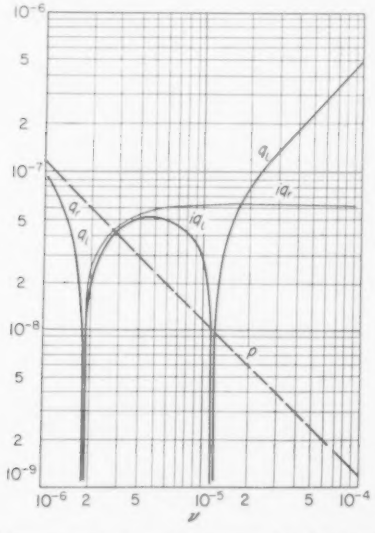

Fig. 7. p and q as functions of frequency ν (iq_r and iq_l means the absolute value of imaginary q_r and q_l).

The values of p and q as functions of ν are shown in Figure 7 for a range of ν from 10^{-4} to 10^{-6} when k is zero. For a range $\nu_{2j} > \nu > \nu_{1j}$, in which q is imaginary, the amplitudes of δ_1– and δ_2–waves, $A_1 \exp(i\delta_{1x})$ and $A_2 \exp(i\delta_{2x})$, which are considered as the reflected waves at the coasts, decrease very rapidly as apart from the coasts except for $\nu \approx \nu_{1j}$ or $\nu \approx \nu_{2j}$. This was defined as a subcritical range (ICHIYE, 1958b). The dominant terms of δ_1– and δ_2–waves in this range become

$$\begin{aligned} A_1\, e^{i\delta_1 x} &\approx \alpha\, R_{j0}\, (p + q_2\, i)^{-1} \exp\{(ip - q_2)\, x\} \\ A_2\, e^{i\delta_2 x} &\approx \alpha\, R_{j0}\, (p - q_2\, i)^{-1} \exp\{i\alpha - (ip + q_2)\, (a - x)\} \end{aligned} \tag{44}$$

because $aq_2 \gg 1$ for a width $a = 6 \times 10^3$ km (the Atlantic). These equations indicate to us that the amplitudes of δ_1– and δ_2–waves decrease very rapidly as apart from the coasts, owing to the highly dispersive nature of the waves.

On the other hand, in the ranges of ν for which q is real, the amplitudes A_1 and A_2 given by (42) oscillate greatly except near $\nu = \nu_{1j}$ or $\nu = \nu_{2j}$ due to the condition $aq > 1$. This phenomenon was discussed by MUNK (1957) for the Rossby waves and this contradiction is avoided by the introduction of frictional force (ICHIYE, 1958b). It can be proved easily that $p_2 \geqslant 0$, $q_2 \geqslant 0$ for all values of ν considered here, when k is finite. The values of p_2 and q_2 for $k = 10^{-6}$ are shown in Fig. 8.

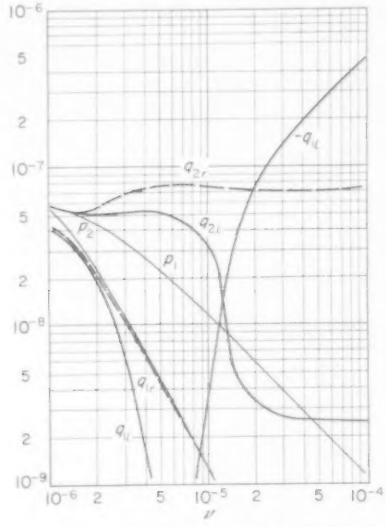

Fig. 8. Real and imaginary parts of p and q for a frictional coefficient $k = 10^{-6}$ ($p = p_1 + ip_2$, $q = q_{1j} + iq_{2j}$, $j = r$ and l).

Since $aq_2 > 1$ for a range of ν considered here except for $\nu > 1{\cdot}5 \times 10^{-5}$ in a baroclinic mode, the predominant terms of δ_2–waves become

$$A_1 e^{i\delta_1 x} = \alpha e^{i\delta_1 x} R_{j0} (p + q)^{-1} \tag{44a}$$

On the other hand, those of δ_2–waves are given by

$$A_2 e^{i\delta_2 x} \approx \alpha (e^{i\gamma a} - e^{i\delta_1 a}) e^{i\delta_2 (x-a)} (p - q)^{-1} R_{j0} \tag{44b}$$

for $\nu < 2 \times 10^{-6}$ and

$$A_2 e^{i\delta_2 x} \approx \alpha \exp\{i\gamma a + i\delta_2 (x - a)\} (p - q)^{-1} R_{j0} \tag{44c}$$

for $\nu > 2 \times 10^{-6}$, in which $q_2 \geqslant p_2$.

From (44a) the amplitudes of δ_1–waves are proportional to $\exp\{-(p_2 + q_2) x\}$, so that they decrease rapidly as apart from the western boundary $x = 0$, since $aq_2 > 1$. However, in the range $\nu < 2 \times 10^{-6}$, in which (44b) is valid, the δ_2–waves decrease more gradually as apart from the eastern coast at $x = a$, they are proportional to $\exp\{(p_2 - q_2)(a - x)\}$ with $p_2 \approx q_2$. This corresponds to the westward intensification of the Rossby waves, which was discussed in detail elsewhere (ICHIYE, 1958b). In the range of $\nu > 2 \times 10^{-6}$, again the δ_2–waves are finite only very close to the eastern coast, since $q_2 > p_2$.

Vol 6 1959

In the exceptional range of $\nu > 1{\cdot}5 \times 10^{-5}$ in a baroclinic mode, which corresponds to the gravity waves, q_l becomes real without a friction (Fig. 7). Even with a friction, the decrease in the amplitudes of the reflected waves is not so conspicuous as for the range of smaller ν, because $aq_2 \approx 1{\cdot}5$ for $\nu > 2 \times 10^{-5}$, $k = 10^{-6}$, and $a = 6 \times 10^3$ km in (44a) and (44c) which are valid for δ_1– and δ_2–waves in this range of ν. Also, the amplitudes of δ_1– and δ_2–waves at the coasts decrease as their frequency increases for $\nu > 2 \times 10^{-5}$, because

$$\left|\alpha (p + q)^{-1}\right| \approx \left|\alpha (p - q)^{-1}\right| \approx \alpha \left|q\right|^{-1} \approx \alpha\, c_l\, \nu^{-1}$$

whereas in ordinary gravity waves the amplitudes of the reflected waves are independent of their frequency.

The proper oscillation of the ocean as a whole should be mentioned, because standing planetary waves in a meridional ocean have periods of several days in the Atlantic (about 5 days at N30° and S30°) and in the Pacific (about 3 to 3·5 days at N30° and S30°) (ARONS and STOMMEL, 1956). This period seems to have some relation to the observed period in the Equatorial Pacific. This is not true in the present system, because the period of standing waves in the equatorial region is much longer than that for the Rossby wave range. To obtain the frequency of proper oscillation of the ocean, the solution of (21) without an external force

$$R_j = A_1 \exp\left\{i\,(\omega t + \delta_1 x)\right\} + A_2 \exp\left\{i\,(\omega t + \delta_2 x)\right\}$$

should satisfy the boundary conditions (40). This gives

$$e^{i\delta_1 a} - e^{i\delta_2 a} = 0 \tag{45}$$

This is also satisfied by

$$a\,\delta_{2n} = a\,\delta_{1n} + 2n\,\pi, \quad n = 1, 2, 3 \ldots . \tag{46}$$

where δ_{1n} and δ_{2n} are the roots of the quadratic equation

$$\delta^2 - \delta\beta\omega^{-1} + l^2 - c_j^{-2}\,\omega^2 = 0 \tag{47}$$

From (46) and (47) we have the relation for ω_n of the n-th mode of the proper oscillation such as

$$\omega_n^4 - c_j^2\,(l^2 + a_n^2)\,\omega^2 - \tfrac{1}{4}\,c_j^2\,\beta^2 = 0 \tag{48}$$

with

$$a_n^2 = n^2\,\pi^2\,a^{-2} \tag{49}$$

We can get two ranges of proper oscillation, one in Rossby waves and the other in gravity waves. In the former, the proper frequency is expressed by

$$\omega_{1,n} = \tfrac{1}{2}\,\beta\,(1 + a_n^2\,l^{-2})^{-\frac{1}{2}}\,l^{-1} \tag{50}$$

However, $a_n^2\,l^{-2}$ is very small for a small number n, since, with $a = 6000$ km $n = 24$ gives $a_n^2\,l^{-2} = 1$. In other words, many proper frequencies are crowded near the frequency $\omega_{1,0} = \frac{1}{2}\,\beta\,l^{-1}$ corresponding to a period of about 39·7 days. In higher latitudes, where the geostrophic approximation is valid, the period of proper oscillations in the Rossby wave range is several days in the first mode. In the equatorial region, on the other hand, it becomes longer, so that it cannot be associated with the observed dominant period of mean sea-level fluctuation.

The proper frequency of gravity waves which are another root of (48) is given approximately by

$$\omega_{2,n} = c_j l (1 + a_n^2 l^{-2})^{\frac{1}{2}} \tag{51}$$

In this range of ω many proper frequencies are crowded near $\omega_{2,0} = c_j l$ ($c_r l = 1{\cdot}29 \times 10^{-3}$, $c_l l = 1{\cdot}23 \times 10^{-5}$ or the period in a baroclinic mode in 5·9 days). The crowding of the proper frequencies near $\omega_{1,0}$ and $w_{2,0}$ is the reason for the wild oscillation of the amplitudes of waves in (42) in the Rossby wave and gravity wave range. Particularly, $\omega_{2,0}$ in a baroclinic mode gives a period very close to the observed one, which suggests that even random disturbances may cause the proper oscillation of the ocean of the observed period.

(7) EFFECT OF THE NORTH-SOUTH EXTENT OF THE WIND SYSTEM

The assumption $(\partial/\partial t)^2 \gg f^2$ ceases to be valid at the outer edge of the wind system when the period of the motion exceeds 6·3 days for a value of $l = 2\pi \times 10^{-8}$. For a longer period, we need two systems of the equations, one is (16) (17) near the equator and another is (19) (20) in the sub-tropical zone.

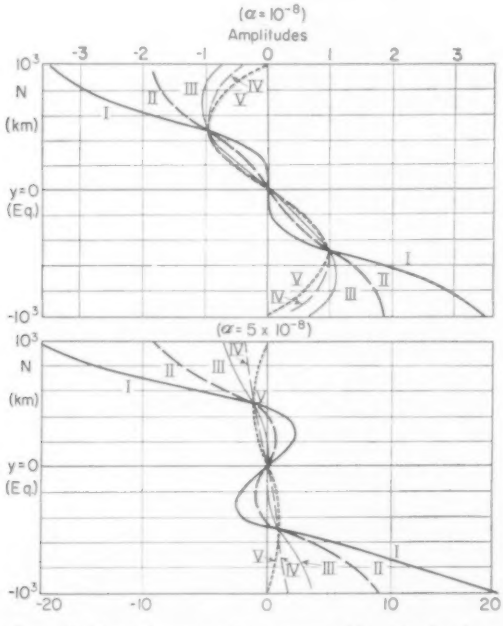

Fig. 9. Distributions of amplitudes of geostrophic waves. (Amplitudes in $\tau_0 \sigma_j l c^{-2} M_j^{-1}$. Curve I for $\nu = 10^{-6}$, II for $\nu = 2 \times 10^{-6}$, III for $\nu = 2 \times 10^{-6}$, IV for $\nu = 5 \times 10^{-6}$, V for $\nu = \infty$).

The latter system can be also separated into normal modes and the equation of R_j becomes

$$\partial (\nabla^2 R_j - \alpha_j^2 \lambda^{-2} R_j)/\partial t + \beta \partial R_j/\partial x = c_j c^{-2} (f \operatorname{curl} \tau + \partial \operatorname{div} \tau/\partial t + \beta \tau_x) \tag{52}$$

Vol 6 1959

In order to know the relation of two systems, a wind stress distribution can be assumed as simple as possible, so that we put

$$\tau_x = 0\,; \quad \tau_y = -\tau_0 \cos ly \cos(\nu t + \alpha x) \tag{53}$$

The solution of (52) becomes

$$R_j = \tau_0\, \sigma_j\, lc^{-2}\, M_j^{-1} (f\alpha\nu^{-1} l^{-1} \cos ly - \sin ly) \cos(\nu t + \alpha x) \tag{54}$$

where

$$M_j = l^2 + \alpha^2 - \alpha\beta\nu^{-1} + \alpha_j^2 \lambda^{-2} \tag{55}$$

In this equation $\sin ly$ term is produced by the divergence of wind stresses and it is similar to the forced motion given by (26) for $D = l\tau_0$, with the only difference of a factor M_j. We can put $f = \beta y$ near the equator in (54) and although this solution is not valid just on the equator, it is applied for the most part near the equator for a range of ν less than 10^{-5}. In Fig. 9 the amplitude distribution is plotted as a function of y at $\alpha = 10^{-8}$ and 5×10^{-8} for $\nu = 10^{-6}$ to 10^{-5}. It is interesting that the distribution of the amplitude of (54) concerning y direction approaches that obtained under the assumption $\mathbf{P}^2 \gg f^2$ as ν increases (Fig. 9). The reason for this condition is that the magnitude of curl-term increases more rapidly than div-term as ν decreases.

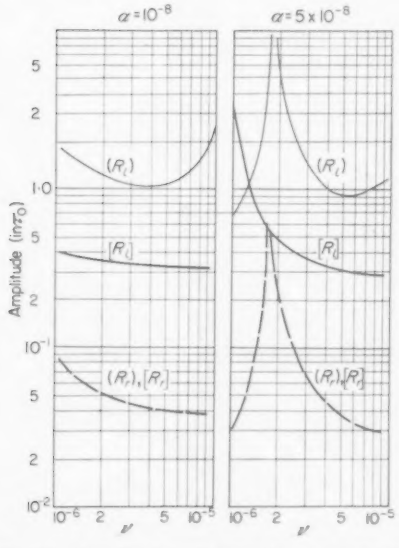

Fig. 10. Amplitudes of waves in an equatorial region (R_j) and waves of geostrophic approximation $[R_j]$.

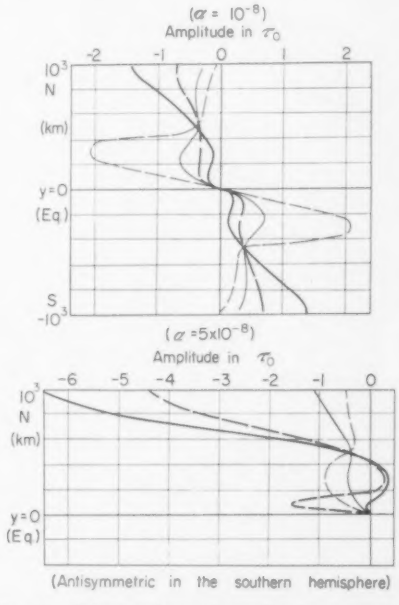

Fig. 11. Distributions of amplitudes corrected by a geostrophic approximation in a subtropical zone for a baroclinic mode.

In order to compare the solution (54) with (26), the amplitude $[R_j] = \tau_0\, l\sigma_j\, c^{-2}\, M_j^{-1}$ should be compared with that of the previous solution (R_j) of (26). In the barotropic mode, $\nu^2 c_r^{-2}$ and $\alpha_j^2 \lambda^{-2}$ are negligibly smaller than other terms, so that $[R_r] \approx (R_r)$. The results of computation of $[R_l]$ for $\alpha = 10^{-8}$ and 5×10^{-8} are shown in Fig. 10, simultaneously with the (R_l) and (R_r) obtained previously.

These curves indicate that $[R_l]$ is generally smaller than (R_l), and that particularly for $\alpha = 5 \times 10^{-8}$ there is no resonant frequency in the curve of $[R_l]$. We cannot explain the predominant period of observed mean sea-level variation from the solution of (54).

As stated earlier, the solution (54) must be applied to the extra-tropical region even in the considered range of ν from 10^{-6} to 10^{-5}, and this solution must be combined with that given by (26) for a comprehensive distribution of the amplitude from the equator to about 1000 km north and south. This distribution is shown in Fig. 11, with one curve based on (54) and the other on (26).

As stated earlier, for a wave number $\alpha = 5 \times 10^{-8}$ the wind stress curl is much larger than divergence, and the amplitude distribution in most regions except very close to the equator follows the pattern for the curl, rather than the divergence. However, near the resonance frequency in a baroclinic mode and also for ν larger than 10^{-5} the equation (26) dominates in most of the area considered.

(8) THE EFFECT OF CURRENTS ON LONG WAVES

In the equatorial ocean, flow steady zonal currents such as the South Equatorial Current and farther north the Equatorial Counter Current. These should have some influences on the long waves discussed in this paper. The complete analysis of the effect of currents on the long waves would lead to non-linear equations difficult to solve. Moreover, information on the detailed structures of the currents is necessary for a rigorous mathematical analysis.

Therefore, to emphasize the understanding of the basic process, we will use as simple a mathematical model as possible. The basic currents are assumed to be constant in the upper and lower layers. This assumption yields linear wave equations, which do not indicate an interaction between the waves and currents. Such a linearization is reasonable, because the waves of a relatively high frequency considered in this paper seem to have no effect on currents.

The equations of motion and continuity are almost similar to those of (1) to (6), under such an assumption, except the operator $\mathbf{P}$ and $\mathbf{P}'$, which must be replaced by

$$\mathbf{Q} = (\partial/\partial t) + U(\partial/\partial x), \quad Q' = (\partial/\partial t) + U'(\partial/\partial x) \tag{56}$$

in which U and U' are basic currents in x-direction assumed to be constant. Then we obtain equations corresponding to (8), (9), (16), (17) and (19), (20) by replacing $\mathbf{P}$, $\mathbf{P}'$ with $\mathbf{Q}$, $\mathbf{Q}'$.

For a moving wave system such as expressed by $\exp\{i(\nu t + \alpha x)\}$,

$$\mathbf{Q} = i(\nu + \alpha U), \quad \mathbf{Q}' = i(\nu + \alpha U').$$

Therefore, when a frequency satisfies $\nu = -\alpha U$ or $\nu = -\alpha U'$, $\mathbf{Q}$ or $\mathbf{Q}'$ vanishes. In other words, when the waves and the currents travel in the same direction, there is a critical phase velocity for the waves, which makes the assumption $\mathbf{Q}^2$, $(\mathbf{Q}')^2 > f^2$ cease to be valid.

If this critical velocity occurs in the upper layer, then the equation (16) must be replaced by that derived from the equation (19) inserting $\mathbf{P} = 0$ for the region not so close to the equator, and near the equator $\mathbf{P} = k$ with the assumption that $k^2 \gg f^2$.

We cannot solve the final equations (16), (17) or (19), (20) by separation into normal modes, because of the characteristic that basic currents U and U' differ from each

Vol 6 1959

other, particularly in the equatorial region, as indicated by many observations (SVERDRUP *et al.*, 1942).

When we put $\eta + \eta' = \phi$, $\xi\eta + \eta' = \psi$, the equations for ϕ and ψ are obtained from the modified relations for (16) and (17) such as

$$(\mathbf{SS}' - \xi\,\mathbf{MM}')\,\phi = \mathbf{S}'F, \quad (\mathbf{SS}' - \xi\,\mathbf{MM}')\,\psi = -\xi\,\mathbf{MM}'F \tag{57}$$

in which $\mathbf{S}$, $\mathbf{S}'$ and $\mathbf{M}$, $\mathbf{M}'$ are linear operators expressed by

$$\mathbf{S} = \mathbf{Q}\,(\nabla^2 - c_1^{-2}\,\mathbf{Q}\,\partial/\partial t) + \beta\,\partial/\partial x, \quad \mathbf{S}' = \mathbf{Q}'\,(\nabla^2 - c_2^{-2}\,\mathbf{Q}'\,\partial/\partial t) + \beta\,\partial/\partial x \tag{58}$$

$$\mathbf{M} = c_1^{-2}\,\mathbf{Q}^2, \quad \mathbf{M}' = c_2^{-2}\,(\mathbf{Q}')^2\,; \quad c_1 = c\,(1-\xi)^{\frac{1}{2}}, \quad c_2 = c'\,(1-\xi)^{\frac{1}{2}} \tag{59}$$

with an abbreviation

$$F = c^{-2}\,(\mathbf{Q}\,\mathrm{div}\,\tau + \beta\tau_x).$$

In a numerical computation of the solution (57) we can put $U' = 0$ without any loss of generality of the theory. In fact, in the equatorial region, currents are very shallow and limited only to the upper layer (SVERDRUP *et al.*, 1942). Since the South Equatorial Current flows westward, we should put $\alpha U = -\kappa\nu$, with a positive parameter κ. Then the left and right sides of the equation (57) are written as

$$\begin{aligned}\mathbf{SS}' &= -\left\{\nu\,(1-\kappa)\,(\nu^2\,c_1^{-2}\,\overline{1-\kappa} - l^2 - \alpha^2) + \alpha\beta\right\}\left\{\nu\,(\nu^2\,c_2^{-2} - l^2 - \alpha^2) + \alpha\beta\right\}\\ \mathbf{MM}' &= -\nu^4\,(1-\kappa)^4\,c_1^{-2}\,c_2^{-2}\end{aligned} \tag{60}$$

and

$$\begin{aligned}\mathbf{S}'F &= -\nu^2\,(1-\kappa)\,c^{-2}\,(\nu^2\,c_2^{-2} - l^2 - \alpha^2 + \alpha\beta\nu^{-1})\,\mathrm{div}\,\tau\\ \mathbf{M}'F &= \nu^4\,(1-\kappa)\,c^{-2}\,c_2^{-2}\,\mathrm{div}\,\tau\end{aligned} \tag{61}$$

From these equations the values of ϕ and ψ are computed as functions of ν for a wave number $\alpha = 5 \times 10^{-8}$. Particularly near the critical frequency $\nu = -\alpha U$, (57) is not valid and instead the next equations must be used.

$$(\mathbf{K}\,\mathbf{S}' - \xi\,\mathbf{N}\,\mathbf{M}')\,\phi = \mathbf{S}'F', \quad (\mathbf{K}\,\mathbf{S}' - \xi\,\mathbf{N}\,\mathbf{M}')\,\psi = -\xi\,\mathbf{M}'F' \tag{57'}$$

where it is assumed that a friction force balances the pressure gradient and wind stresses. The operators $\mathbf{K}$ and $\mathbf{N}$ are defined as

$$\mathbf{K} = k\,(\nabla^2 - k\,c_1^{-2}\,\partial/\partial t) + \beta\,\partial/\partial x \tag{58'}$$

$$\mathbf{N} = k^2\,c_1^{-2}\,(\partial/\partial t) \tag{59'}$$

with

$$F' = c^{-2}\,k\,\mathrm{div}\,\tau$$

For numerical computation, the velocity U is taken to be equal to -10^2 (cm). Then near the critical frequency $\nu = 5 \times 10^{-6}$, (57′) becomes simply

$$\begin{aligned}\phi &= k\,\mathrm{div}\,\tau\,c^{-2}\left\{\alpha\beta + ik\,(l^2+\alpha^2)\right\}^{-1}\\ \psi &= \xi\nu^2\,\phi\,c_2^{-2}\,(l^2 + \alpha^2 - \alpha\beta\nu^{-1})^{-1}\end{aligned} \tag{62}$$

due to the condition that

$$|\mathbf{N}|\;|\mathbf{K}|^{-1} = 0{\cdot}94 \times 10^{-3}, \quad |\mathbf{M}'|\;|\mathbf{S}'|^{-1} = 0{\cdot}73 \times 10^{-3}$$

for $k = 10^{-6}$.

These numerical results are shown in Figs. 12 (a), 12 (b) and 13. The function ϕ represents the surface elevation and the deviation of the intersurface η' is computed from the relation $(\psi - \xi\phi)(1 - \xi)^{-1}$. On the other hand, the characteristic velocity in the upper and lower layer v_g, v_g' defined in the Section 3 is proportional to $\phi\nu^{-1}$ and $\psi\nu^{-1}$. Since the equations for ϕ and ψ have the 6th order time-derivative, rather

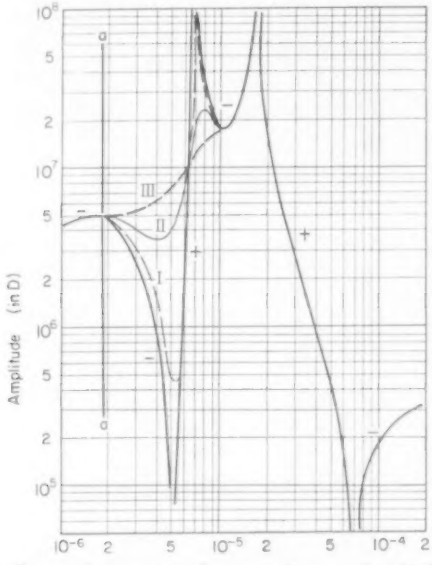

Fig. 12 (a). Amplitudes ϕ of waves in presence of currents for $\alpha = 5 \times 10^{-8}$. (Thick lines: no correction near a critical frequency ω_1'. I : corrected curve for $k = 10^{-7}$, II: for $k = 10^{-6}$, III: for $k = 10^{-5}$. Plus or minus sign means the positive or negative value. The portion indicated with *a* is shown in detail in Fig. 12 (b).

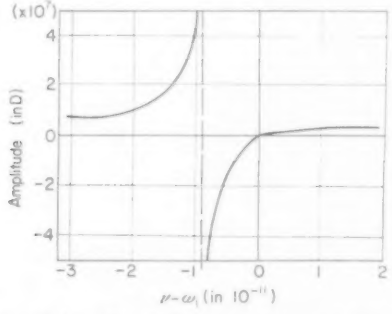

Fig. 12 (b). Detailed curve of ϕ near $\nu = \omega_1$.

than the third order for normal modes R_j, the number of poles of ϕ and ψ as functions of ν increases from the number in the solution of normal modes. Particularly near the zero of ϕ in the Rossby wave range corresponding to the zero $\nu = \omega_1$ of $\mathbf{S}'F$ in (61), there is a pole of ϕ. Near $\nu = \omega_1$, ϕ is expressed as

$$\phi = (1 - \kappa)\,(\nu - \omega_1)\,c^{-2}\,|\mathbf{S}|^{-1}\,(\nu - \omega_1 + \vartheta\,\omega_1)^{-1}.$$

Vol 6 1959

In this equation, $|\mathbf{S}|$ is the absolute value of an operator substituted with ν and α into $\partial/\partial t$ and $\partial/\partial x$ and

$$|\mathbf{S}| = 1{\cdot}81 \times 10^{-14}, \quad \omega_1 \approx \alpha\beta\,(l^2 + \alpha^2)^{-1} = 1{\cdot}79 \times 10^{-6}$$

$$\vartheta = \nu^4\,(1 - \kappa)^2\,\xi\,c^{-2}\,c_2{}^{-2}\,|\mathbf{S}|^{-1} = 0{\cdot}82 \times 10^{-5}.$$

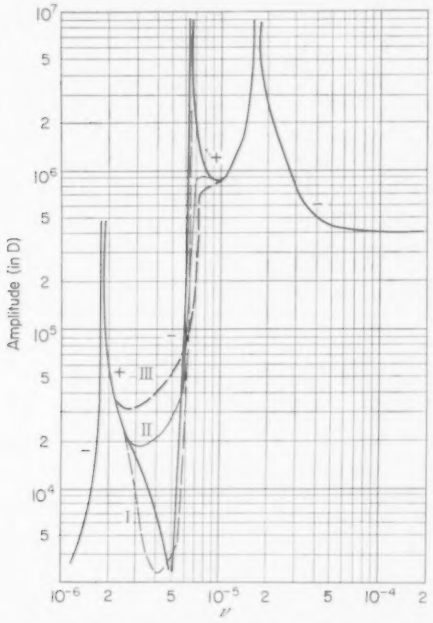

Fig. 13. Amplitude ψ of waves in presence of currents. (See legend to Fig. 12.)

Therefore, the amplitude of resonance is confined, within very narrow band of $\nu = \omega_1 \pm 10^{-11}$ (Fig. 12a), so that this resonant frequency almost vanishes, as there is almost no increase in amplitude very close to ω_1. On the other hand, another resonant frequency appears near $\omega_1{}' = 6{\cdot}8 \times 10^{-6}$ which is also in the Rossby wave range, but corresponds to the zero of $|\mathbf{S}|$ in (6). However, this resonance belongs to a range of ν in which the assumption $\mathbf{Q}^2 \gg f^2$ is not valid. The corrected curves in this range from the equation (62) are also shown in the figure, and if a frictional force is assumed to be responsible for balancing the pressure gradient even near ω'_1, then the resonance at $\omega_1{}'$ almost vanishes, for a frictional coefficient k larger than 10^{-6}.

On the other hand, the resonant frequency in the gravity wave range shifts upwards to a higher side due to the effect of currents, since ω_2 becomes equal to $1{\cdot}78 \times 10^{-5}$ or it corresponds to a period of 4·02 days instead of 4·9 days in the case of no current. This affords a better explanation of the observed period in the Pacific islands.

The existence of currents has greater influence on the Rossby wave range than the gravity wave range, but the results for higher frequency waves associated with the observed sea-level variation are not changed.

(9) SUMMARY OF THE THEORY AND CONCLUDING REMARKS

The present theory for long waves in a stratified ocean near the equator is summarized as follows :

(1) The vorticity equation for a stratified ocean near the equator leads to wave equations concerning deviations of sea-level and intersurface level, with the assumption that the square of the frequency is negligible compared with that of the Coriolis' parameter. These equations differ from those in mid-latitudes because the external force is a divergence of the wind stresses rather than curl and also the divergence term is associated with gravity waves rather than inertia gravity waves.

(2) The equations are solved by a method of separation into normal modes. The free waves obtained by this method are gravity waves of higher frequency and Rossby waves of lower frequency. The separation of these waves is almost complete in a baroclinic and barotropic mode.

(3) The forced waves caused by a moving wind system have a resonance at a higher frequency of the gravity wave range. For a westward moving wind system corresponding to easterly waves in the tropical region there is a resonance with a frequency in the Rossby wave range. A resonant period of about five days occurs in the baroclinic gravity wave range for this wind system. This period agrees well with the observed predominant 4-day period of the sea-level variation in the Pacific islands near the equator.

(4) When a moving disturbance appears suddenly, free waves and forced waves are initiated simultaneously. Free waves with three components, one Rossby waves moving westward, the other two gravity waves moving both to the east and to the west. At the resonant frequency the forced wave and one of the free waves cancel out. Near this frequency interference of these two waves causes relatively high amplitudes. Therefore, the predominant waves of four days period might be caused by such a sudden disturbance.

(5) The effect of friction in reducing the amplitudes with time in Rossby waves is almost twice that found in gravity waves.

(6) At the coast, the Rossby and gravity waves are reflected. However, the waves of a frequency intermediate between the two are highly dispersive, and are scarcely dispersive, and are scarcely reflected. The effect of the coast is thus limited to the immediate vicinity of the coast.

(7) Baroclinic gravity waves have proper oscillations with a period of about 6 days between an east and a west coast.

(8) The equations obtained with the assumption of (1) are applicable to a region very close to the equator and also for a frequency higher than the Rossby wave range. The solution must be modified for a wind system with a wider north-south extension or with a lower frequency, in which the geostrophic relation holds. Near the resonant frequency in baroclinic gravity waves, they are valid in a wider area corresponding to the whole easterly waves.

(9) The existence of currents, particularly westward-flowing ones such as the South Equatorial Current, have a great influence on waves of lower frequencies, and in fact such westward currents suppress the resonance in the Rossby wave range if a frictional force is present. On the other hand, they have almost no effect on the waves of a higher frequency, and they reduce the resonant period of baroclinic gravity waves to 4 days, which are closer to the observed period of dominant oscillations.

Vol
6
1959

As stated earlier, the present work only clarifies the basic mechanism of long waves near the equator. Data for the waves considered here is quite meagre, except for fluctuations of mean sea-level in the equatorial islands (GROVES, 1956) as mentioned earlier. Therefore, it is almost impossible to discuss the details for the application of this theory. Except for comparison of the predominant period, it is only possible to estimate theoretically the order of amplitudes of mean sea-level fluctuations caused by easterlies. The observed amplitude of sea-level at Canton Island is about 5 to 10 cm (GROVES, 1956). On the other hand, divergence of the wind (not wind stresses) in the easterly waves is from -7 to 10^{-6} to 3×10^{-6} (PALMER, 1952) with wind speed of about 5 to 20 knots. By taking 0·2 as the ratio of wind stresses to the wind speed, corresponding to the average wind stress factor between 4 m/s to 10 m/s, an amplitude of the wind stress divergence becomes about 10^{-6}. From the curve of $\eta + \eta'$ in Fig. 3, or that of ϕ in Fig. 12, for a wave number $\alpha = 5 \times 10^{-8}$, which corresponds to that of easterly waves, the amplitude of sea level $\eta + \eta'$ has the order of more than 5 cm within $\pm$ 0·3 days of the resonant period (or within $\pm 10^{-6}$ of the resonant frequency). This order of magnitude of sea-level oscillation seems to agree very well with the observed one.

The characteristic of the resonance discussed here as an explanation for the observed predominant oscillation is that it is in a baroclinic mode. Therefore, some hydrographic observations must be made to confirm the theory, and it is hoped that some day such observations will be obtained.

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A quantitative study of phytoplankton from the equatorial Pacific

GRETHE RYTTER HASLE

(*Received* 10 *February* 1959)

Abstract—The hydrography and abundance of phytoplankton were studied at three stations located in the equatorial Pacific, March 1957. Water bottle samples preserved with neutralized formalin were used for the phytoplankton study, which included the smaller forms of diatoms, dinoflagellates and coccolithophorids. Difficulties in obtaining reliable estimates in the case of poor populations were overcome by a simple concentration technique.

The quantity of phytoplankton, computed as number of cells under 0·1 m^2, was found to be much the same as the maximal figure for the Sargasso Sea and the amount obtained before and after the time of maximal population in the antarctic and subantarctic Pacific.

In the equatorial Pacific phytoplankton abundance was restricted to the upper 50 to 100 m. Maxima in the numbers of diatoms were found nearer the surface than those of dinoflagellates, while the coccolithophorids showed different patterns in their vertical distribution. " Olivgrüne Zellen " (HENTSCHEL) were present at the greatest sampling depths.

A pennate diatom species, athecate dinoflagellates and certain small coccolithophorids were the most numerous forms observed.

Circumstantial evidence of seasonal variation is discussed.

INTRODUCTION

THIS study is based on material collected at three stations in the central equatorial Pacific during cruise 38 of the United States Fish and Wildlife Service vessel *Hugh M. Smith*, in March 1957 (Fig. 1).

The main purpose of the study is to compare quantitatively and qualitatively the phytoplankton from the subantarctic and the equatorial region of the Pacific Ocean*.

The material also affords opportunity for studying the vertical distribution of phytoplankton in the tropics, to test the general validity of the theory of Schimper (CHUN, 1900) and LOHMANN (1920) regarding the " shadow flora " and HENTSCHEL (1928) and SCHILLER (1931) concerning " Olivgrüne Zellen."

PREVIOUS INVESTIGATIONS IN THE AREA

Phytoplankton

With the exception of GRAHAM (1941) who dealt with dry weight of plankton hauls (both phyto- and zooplankton), the literature contains no information about quantitative investigations of phytoplankton from an oceanic locality in the Pacific Ocean. MARSHALL's study (1933) from the Great Barrier Reef can be compared with the present investigation regarding the collecting method, but the locality is not oceanic. Quantitative investigations from the western Pacific (AIKAWA 1936, MOTODA 1941, JOHNSON 1954) also deal with phytoplankton of lagoons, bays and coastal waters, and are all based on material collected by nets.

*The author is studying phytoplankton material from the Pacific part of the Antarctic and Subantarctic collected by the *Brategg*-Expedition, the Norwegian Antarctic Expedition 1947-1948. The results will be published in the series of scientific results of this expedition.

Vol 6 1959

Allied investigations

In other fields of oceanography the equatorial part of the Pacific Ocean has been rather intensively studied, especially during the last ten years by the staffs of the Pacific Oceanic Fishery Investigations (POFI) and the University of Hawaii, both at Honolulu.

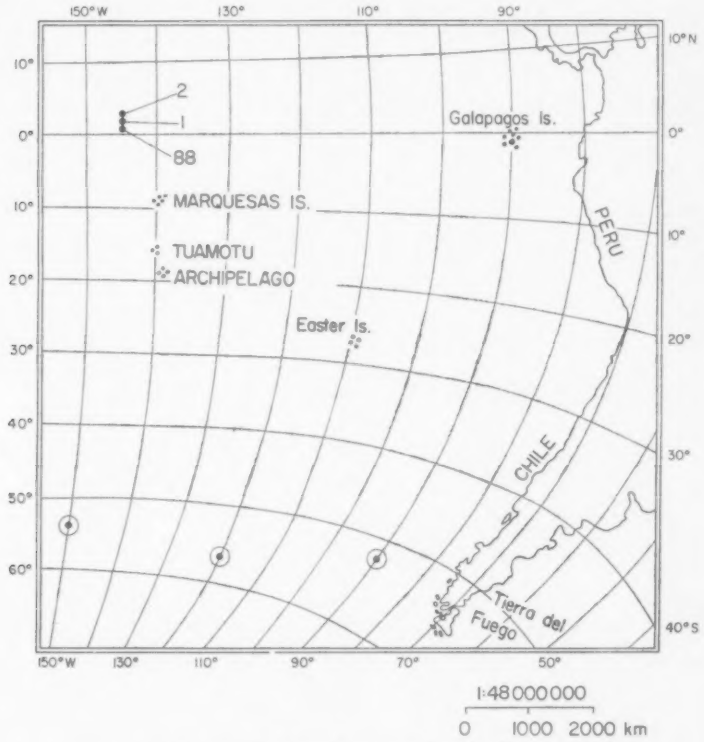

Fig. 1. The south-eastern Pacific Ocean.
⊙ The northernmost stations of the *Brategg*-Expedition, 1947-48.
● The phytoplankton stations of *Hugh M. Smith*, March 1957.

The hydrography has been most thoroughly investigated (for references see STROUP and AUSTIN, 1955). The density of tuna larvae has been studied by SETTE (1955), the abundance of zooplankton by KING and DEMOND (1953) and the vertical distribution of zooplankton by HIDA and KING (1955). More recently, chlorophyll observations were made (KING *et al.* 1957), but the data have not yet been fully studied. Measurements of the photosynthetic carbon fixation by the ^{14}C method were made on the *Galathea*-Expedition (STEEMANN NIELSEN, 1957) and also by DOTY *et al.* (1956). Further data are published by KING *et al* (1957) and AUSTIN (1957). The particle distribution in the equatorial Pacific was studied by JERLOV (1953) on the Swedish Deep-Sea Expedition 1947-1948 " in order to find evidence for water movements " (l.c. p. 78). BURT (1958) measured light transmission in the eastern tropical Pacific. Secchi disk readings from the central equatorial Pacific may be found in KING *et al.* (1957) and AUSTIN (1957).

MATERIAL

The phytoplankton samples were collected with water bottles from eight depths at three stations, the greatest sampling depth varying from 267 to 434 m. The samples were preserved unconcentrated with 1 ml neutralized 40 per cent formalin in 250 ml plastic bottles. The temperature, salinity, and oxygen and phosphate contents were determined.

The positions and dates of the stations were :

Station	Latitude	Longitude	Date	Time*
88	00° 01′N	145° 02′W	20.3.57	17:18–18:07
1	00° 58·6′N	145° 05·2′W	21.3.57	04:22–04:48
2	02° 00·7′N	145° 05·6′W	21.3.57	11:56–12:22

All three stations are situated in the South Equatorial Current. The boundary between this and the Equatorial Countercurrent has been reported to lie at 5°N (AUSTIN *et al.* 1956) or between 2·5° and 5°N (JERLOV, 1956).

HYDROGRAPHY

The hydrographical data are presented in the form used by MONTGOMERY (1954) (Fig. 2).

The stations show the same pattern of temperature, salinity and oxygen distribution as usually found in this region. The surface temperatures are about 27°C. The thermocline is very pronounced at depths between 100 and 150 m and, at depths of about 200 m, the temperature is about 13°C. Salinity is somewhat lower in the upper 50 m than in the next 100 m (particularly at Stations 88 and 2) and then begins to decrease at about 150 m. The salinity at the three stations, from the surface down to about 300 m, ranges from 34·70 to 35·30‰. The oxygen content is about 4·80 ml/l at the surface and decreases with depth. At Station 88 (the station nearest the equator) hydrographical observations were taken down to 1275 m. This station has an oxygen minimum at 434 m with 0·86 ml/l. At 1275 m the oxygen content is 2 ml/l. Values between 0·5 and 1·0 ml/l at 400-500 m have been recorded in this area (e.g. AUSTIN, 1957), although the well known oxygen minimum in the Pacific usually is recorded from positions further south and north (RICHARDS, 1957 p. 224 ff.). The percentage saturation was computed after RICHARDS and CORWIN (1956) and was greater than 100 per cent down to about 50 m, the greatest value being 108 per cent. The phosphate content at the three stations differs. Only the station richest in phosphate, situated nearest to the equator (Station 88), had a ' normal ' phosphate content compared with other observations by POFI (e.g. CROMWELL and AUSTIN, 1954, STROUP, 1954, AUSTIN, 1954, 1957 and KING *et al.* 1957) with a surface value of 0·51 μg-atoms/l. At Station 1 the corresponding value was 0·17 μg-atoms/l. and at the northernmost station 0·24 μg-atoms/l. Such low values are usually found only at higher latitudes. Along the equator, on the other hand, a divergence followed by a band of upwelling and mixing is regularly found, probably first observed by the *Carnegie*-Expedition (cf. SVERDRUP *et al.* 1946 p. 711), later confirmed by JERLOV (1956) and the cruises of POFI (e.g. CROMWELL, 1953 and STROUP and AUSTIN, 1955).

*Time corresponding to + 10 zone time.

Vol 6 1959

An upwelling area in these latitudes is usually characterized by low salinity, temperature and oxygen content and high phosphate values. In spite of the higher phosphate content at Station 88, as discussed above, the present hydrographical data do not

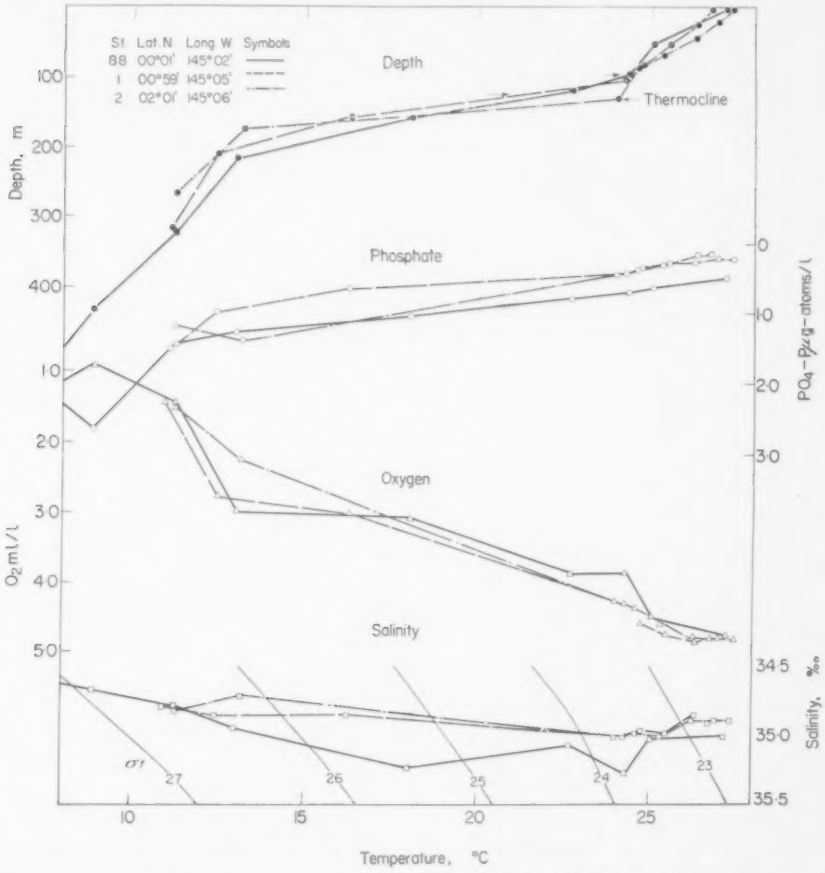

Fig. 2. Temperature, phosphate content, oxygen content and salinity at Stations 88, 1 and 2.

support unequivocally the assumption that this station is situated in an upwelling area. A possible explanation for the higher phosphate value might be the occurrence of a meridional transport of upwelled water originating south of this area.

PHYTOPLANKTON

Method of examination

Since UTERMÖHL (1931) introduced the inverted microscope in quantitative phytoplankton investigations, most of the European scientists working with these problems have considered his sedimentation method superior to other quantitative techniques.

The concentration of organisms in small water volumes (1-100 ml) might be very inadequate in a quantitative study of barren waters. UTERMÖHL (1927) discussed

this in connexion with the sparse occurrence of certain smaller coccolithophorids. In the older Kolkwitz cell technique, the original UTERMÖHL technique and its recent modification (cf. UTERMÖHL, 1958), the organisms of the larger sub-samples are usually sedimented on a rather large area (the bottom of the counting cylinder). Because of the time involved, the examination of such sub-samples has to be made at a rather low magnification, and rare phytoplankton forms of small size and fragile appearance are not adequately recorded.

The best way to solve this problem might be to concentrate the organisms in a smaller volume. This could be done by allowing the organisms to settle and subsequently pipetting off most of the supernatant liquid. The concentrated sample may then be examined in a smaller counting cylinder at a greater magnification.

BERNARD and FAGE (1936) tried in this way to adapt the UTERMÖHL technique for their study of the scarce phytoplankton of the Mediterranean. The loss of cells caused by the concentration was found to be about 20 per cent, a relatively small error compared with that involved in the method itself. However, they did not recommend this modification because other particles in the sea-water (regularly more numerous than the organisms) impede the microscopical examination.

HALLDAL (1953) concentrated one litre samples and found that the loss of coccolithophorids by concentration was very large. As the concentrate was then diluted to 100 ml with distilled water, it is probable that the coccoliths were dissolved during such treatment; this was later demonstrated in our laboratory.

Examination of concentrated samples in counting chambers (mostly Sedgwick-Rafter cells) is used in American phytoplankton investigations. The loss by concentration was tested by CONOVER (1956) by filtering the supernatant liquid through a millipore filter and by counting the organisms retained by the filter after clearing the latter. The loss of organisms was extremely low, only 0·2 per cent.

When examining the present material the following procedure was used : The preserved, unconcentrated sample was agitated thoroughly, poured into glass cylinders of 2 and 50 ml capacity, and left to sediment over night. Counts were then made with an inverted microscope. The organisms were identified to species whenever possible (the 2 ml cylinder was examined at a magnification of about $\times 200$, the 50 ml cylinder of about $\times 60$). In many cases it was necessary to transfer organisms by means of a micropipette (HALLER NIELSEN, 1950) to a regular compound microscope for identification.

Before counting, some of the poorest samples, namely those from the lower depths, were concentrated in small tubes by sedimentation, the supernatant water was pipetted off, and the rest transferred to a 2 ml counting cylinder. If it was necessary to add water to get exactly 2 ml, pipetted sea-water was always used. In this way a magnification of $\times 200$ would be used in the examination of the whole preserved sample of 250 ml.

When increasing the size of the examined aliquot by concentration, two problems are encountered : firstly, how far does the number of species identified depend upon the size of the aliquot, and secondly, to what degree is the number of specimens influenced by the concentration, in other words how great is the loss of cells.

The first problem was discussed by LOHMANN (1920) who tried to find what size aliquots should be centrifuged in order to obtain reliable information regarding the distribution of the coccolithophorids in the tropical Atlantic.

Vol 6 1959

In the present study the numerical data obtained from the concentrated sample and unconcentrated aliquots were compared in five cases (Tables 1 and 2).

Table 1. *Estimates of total population and number of species by examination of different volumes*

Number of ml *examined* *Approx. magnification*		2 (× 200)		50 (× 60)		250 (× 200)	
Station	*Depth* (m)	*Number of species*	*Number of cells per litre*	*Number of species*	*Number of cells per litre*	*Number of species*	*Number of cells per litre*
1	158	2	1000	16	2250	42	3176
1	211	1	1000	0	0	12	535
2	137	7	7000	18	2260	36	4464
2	175	3	2500	8	1220	18	1760
2	267	1	500	6	290	18	655
Sum		14	12,000	48	6020	126	10,590

Table 2. *Estimates of number of cells and number of species by examination of different volumes*

Number of ml *examined* *Approx. magnification*		2 (× 200)		50 (× 60)		250 (× 200)	
Station	*Depth* (m)	*Number of species*	*Number of cells per litre*	*Number of species*	*Number of cells per litre*	*Number of species*	*Number of cells per litre*
				Diatoms			
1	158	0	0	5	320	15	336
1	211	0	0	0	0	3	85
2	137	1	500	7	420	17	528
2	175	1	1000	4	480	6	256
2	267	0	0	2	190	7	65
				Dinoflagellates			
1	158	0	0	2	60	12	628
1	211	1	1000	0	0	4	75
2	137	1	1500	1	40	6	412
2	175	2	1500	1	80	4	564
2	267	1	500	1	20	5	75
				Coccolithophorids			
1	158	2	1000	9	1870	15	2212
1	211	0	0	0	0	5	375
2	137	5	5000	10	1800	13	3524
2	175	0	0	3	660	8	940
2	267	0	0	3	80	6	515

As far as the number of species is concerned, the result of the comparison may be expressed in this way :

1. The total number of identified species of diatoms, dinoflagellates and coccolithophorids will increase by a factor of three when the volume examined increases

from 2 to 50 ml, and again by a factor of three when raising the volume from 50 to 250 ml.

2. The number of diatom species increases by a factor of nine when raising the aliquot from 2 to 50 ml, and by a factor of three when increasing the volume from 50 to 250 ml.

3. The examination of 2 and 50 ml gives the same number of dinoflagellate species, but the increase from 50 to 250 ml is by a factor of six.

4. For the coccolithophorids the increase is three times from 2 to 50 ml and two times from 50 to 250 ml.

These results indicate that an enumeration of diatoms in aliquots of 50 ml may be sufficient to give the broad features of the distribution of this group, but that the same volume may be too small for estimation of dinoflagellate abundance in these rather poor samples. In these waters they are evidently represented by many species and by a few specimens of each species. A weak point in this comparison is the use of a lower magnification for the 50 ml aliquots than for the other two. However, the number of species will probably not be affected by this, because, during the enumeration the higher magnification was usually employed in the identification of the smaller forms. It is also noteworthy that when a species was recorded only in the concentrate of 250 ml, it was present in such low numbers that it could not be expected to occur in a smaller volume.

Table 3. *Estimates of number of cells by examination of* 50 ml *sub-samples and concentrates of* 250 ml

Number of ml *examined* *Approx. magnification*	50 (× 60)	250 (× 200)
Anthosphaera sp.		
Station 1, 158 m	860	496
Station 2, 137 m	80	96
Station 2, 175 m	580	528
Cyclococcolithus cf. *fragilis*		
Station 2, 137 m	400	272
Cyclococcolithus leptoporus		
Station 2, 137 m	480	520
Thorosphaera flabellata		
Station 1, 158 m	660	708
Station 2, 137 m	360	256
Syracosphaera sp. (?)		
Station 1, 158 m	120	200

The data given in Table 1 demonstrate how quantitative differences of poor samples are smoothed out when examining rather small volumes (e.g. 2 ml). The number of cells per litre seems to be much less influenced than the number of species by increasing the examined volume (Table 1). Even if this number increases by a factor of 9 when the examined volume is increased from 2 to 250 ml the average number of cells per litre is about the same. This may be explained by the occasional presence of, for example, one cell of a species in 2 ml (500 cells per litre) although the true abundance might be less.

Vol 6 1959

The discrepancy between the number of cells per litre estimated from the 50 ml aliquots and the 250 ml concentrates may partly be caused by the different magnification used (e.g. Table 2, the coccolithophorids). Therefore, in order to estimate the loss of cells by concentration the number of cells per litre of some easily recognizable coccolithophorids enumerated in the 50 ml aliquot and the 250 ml concentrate will be used (Table 3). The results from the 2 ml aliquots are omitted because these species appeared only occasionally in such a small volume. Based on the data in Table 3, the loss by concentration can be estimated to be less than 10 per cent (7 per cent as a mean of these 8 observations).

Table 4. *Standing stock of diatoms, dinoflagellates and coccolithophorids as* 10^6 *cells under* 0·1 m^2

Equatorial Pacific

	Station 88	*Station* 1	*Station* 2
Lat.	00° 01′N	00° 58′N	02° 01′N
Long.	145° 02′W	145° 05′W	145° 06′W
Date	20 March 57	20 March 57	21 March 57
	Standing stock	Standing stock	Standing stock
	To 434 m : 474	To 318 m : 470	To 267 m : 317
	To 159 m : 447	To 158 m : 452	To 175 m : 304

Subantarctic Pacific

	Station 2	*Station* 32	*Station* 45
Lat.	53° 49′S	54° 32′S	62° 29′S
Long.	89° 59′W	149° 58′W	110° 35′W
Date	13 Dec. 48	21 Jan. 48	3 Feb. 48
Secchi disk depth	23 m	15 m	32 m
	During time of max. population	*During* (?) time of max. population	*After* time of max. population
	Standing stock	Standing stock	Standing stock
	To 200 m: 1356	To 146 m : 4660	To 129 m : 656

Tropical and subtropical Atlantic

	35°N 48°W		0°–10°N	8°S–20°S
Lat.				
Long.				
Date	April–May	Sept.–Nov.	July	May–Sept.
	Standing stock	Standing stock	Standing stock	Standing stock
	To 150 m : 344	To 150 m : 44	To 400 m : 22·6	To 1000 m : 49
	Riley 1957 p. 264		Lohmann 1920 p. 16	Hentschel 1936 *in* Steemann Nielsen 1957, p. 110

Antarctic Pacific

	Station 25	*Station* 41	*Station* 50
Lat.	60° 55′S	65° 14′S	64° 30′S
Long.	150° 28′W	120° 33′W	90° 15′W
Date	17 Jan. 48	30 Jan. 48	15 Feb. 48
Secchi disk depth	32 m	7·5 m	
	Before time of max. population	*During* time of max. population	*After* time of max. population
	Standing stock	Standing stock	Standing stock
	To 150 m : 401	To 100 m : 7047	To 120 m : 20

Therefore, when studying phytoplankton of waters rich in species and poor in specimens and disturbing particles, it seems advisable to concentrate the poorest samples.

The standing stock

It must be stressed that these data relate to one of the more productive areas in oceanic, tropical waters. Observations of the phosphate content and productivity measurements made by the POFI readily indicate that there exists a fertile zone from about 10°S to 10°N. Furthermore, the standing stock sampled during the present investigation may be smaller than at other seasons (see p. 56).

The standing stock is presented as number of cells under 0·1 m^2, computed after BIGELOW *et al.* (1940). Only diatoms, dinoflagellates and coccolithophorids are included in these estimates as other flagellates and algae are usually deformed or destroyed by preservation with formalin.

The data in Table 4 demonstrate clearly that there is no significant difference in the standing stock present at Station 88 and Station 1, while that at Station 2 is approximately 33 per cent lower than at Stations 88 and 1.

If the explanation of higher phosphate content at Station 88 is correct (see p. 41), the relatively great phytoplankton abundance at Station 1 may likewise result from meridional transport of waters rich in phytoplankton. A northward displacement of the zooplankton population was discussed theoretically by CROMWELL (1953) and confirmed by KING and DEMOND (1953). Station 2 appears not to be influenced by this proposed mechanism.

A quantitative comparison with other areas

The data in Table 4 may also be used to compare the standing stocks in the equatorial Pacific with those in the tropical and subtropical Atlantic. As far as the sampling and concentration method are concerned, the observations from the equatorial Pacific and subtropical Atlantic are most easily compared. The material from the Sargasso Sea (RILEY, 1957) was counted in a Sedgwick-Rafter cell after concentration. In the other two cases unpreserved water samples were concentrated by centrifugation and counted on a slide (LOHMANN's (1908) centrifugation method), a method which is now little used because of the great loss of cells during the procedure. The numbers recorded by LOHMANN and HENTSCHEL are therefore minimum values.

It seems justified to conclude that in March the area near the equator in the Pacific had a standing stock of the same order of magnitude as the maximum value recorded in the Sargasso Sea (April-May) and definitely higher than that found in the tropical Atlantic by LOHMANN (1920) and HENTSCHEL (1936). This supports the results of previous biological investigations in this area, characterizing the central equatorial Pacific as a productive region compared with other tropical localities.

The data from the subantarctic and antarctic Pacific were obtained by the same method as those presented here from the equatorial Pacific. The waters of the South Polar region are known as some of the richest of all the oceans. But the phytoplankton abundance depends largely upon the stage of development of the population. Before and after the time of maximal abundance the standing stock is of the same order of magnitude or even less than that found in the equatorial Pacific during March, while the maximal population in the Subantarctic and Antarctic is up to 20 times larger.

As a phytoplankton population consists of cells of different sizes, and as the proportion of larger and smaller cells within a population varies in both space and time,

Vol 6 1959

many planktonologists report the standing stock as cell volume rather than cell number. " Plasma-volume " (LOHMANN, 1908) is a somewhat better indicator of standing stock. As the purpose of comparison between standing stocks at the equator and other tropical localities and those in the Subantarctic and Antarctic is to get an idea of the relative size of the phytoplankton population in the equatorial Pacific, a calculation of plasma volume was not considered necessary. But it must be kept in mind that some of the antarctic diatoms are giants compared to the tropical coccolithophorids. The most numerous species in both areas are also the smallest ones.

Another objection may be raised against the comparison given in Table 4, namely that the greatest sampling depth varies between stations. From Table 4 it is seen that the difference between the standing stock in the upper 150 m and down to the greatest sampling depth is only 5-6 per cent for the stations in the equatorial Pacific where the euphotic zone is deeper than in the Subantarctic and Antarctic. From the vertical distribution diagrams it is also evident that the standing stock beneath 150 m will only slightly influence the total standing stock.

Vertical distribution

In presenting the vertical distribution of the three main groups of phytoplankton of the equatorial Pacific, the diatoms, dinoflagellates and coccolithophorids, LOHMANN's spherical curves (cf. LOHMANN, 1908 p. 193) are used (Fig. 3).

The vertical distribution of all groups shows a well known feature; the decrease of cell number with increasing depth. Some coccolithophorids are an exception to this phenomenon. The vertical distribution, however, differs for the groups, probably according to their particular light requirements.

The decrease in diatom abundance as a whole is especially marked at Station 1, between the surface and 24 m (from 44,000, to 6000 cells per litre). At Station 2 a similar decrease is found between 19 and 43 m, a decrease from approximately 17,000 to 2000. At Station 88, unfortunately, no samples were taken between 0 and 50 m, but it is noteworthy that at the surface there were 70,000 approximately, at 50 m 10,000 and at 95 m only 1400 cells per litre. Diatoms other than the *Nitzschia*-species, which in the surface samples made up more than 90 per cent of the number, showed the same tendency in their vertical distribution. At this time of the year the diatoms were pronounced surface forms. The bulk of the cells were restricted to the upper 25 or 50 m. This finding is somewhat surprising because of its similarity to what has been recorded from coastal waters. It is also in contradiction with the observations from the *Valdivia*-Expedition in the tropical Atlantic (the Equatorial Currents and the Guinea Current) on which Schimper founded his theory of the " shadow flora " (*in* CHUN, 1900 p. 73). According to the latter observations the diatoms *Planktoniella sol* and a *Coscinodiscus* species did not occur above 80 to 100 m. In the material from the equatorial Pacific no such *Coscinodiscus* appeared, neither was *Planktoniella sol* restricted to lower depths. With the method employed during the present investigation much smaller amounts of water are sampled than by the use of a closing-net. The proportion between the estimated numbers for the various levels should be roughly the same regardless of the sampling method used, as all the observed *Planktoniella* cells with the winglike expansion are so large that they will be retained by a net. At all three stations maximum numbers of this species were found in the surface samples. Wingless *Planktoniella sol* cells showed a more irregular

vertical distribution, perhaps partly dependent upon difficulties of identification (because of their small size and similarity to other small centric diatoms). But they too were registered from the surface as well as from depths down to 100 m.

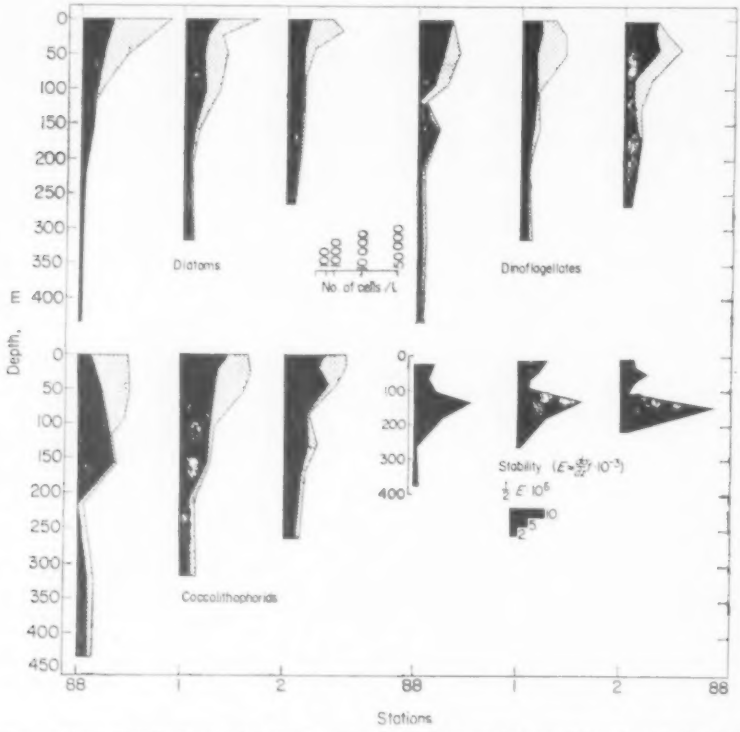

Fig. 3. The vertical distribution of diatoms, dinoflagellates and coccolithophorids by LOHMANN' spherical curves, and of stability. The entire curve presents the total group, the dark filled portion the group omitting some surface forms.

A significant drop in cell numbers of dinoflagellates occurs deeper in the water column than for diatoms. Below 50 to 100 m they do not seem to thrive. Their maximum abundance was found at 50 m at all three stations. This is more pronounced for the gymnodinians than for the other dinoflagellates. The maximum for gymnodinians occurred at 50 m at all the three stations, although the stations were sampled at different times of day (Station 88 in the afternoon, Station 1 at 4.0 a.m., Station 2 at noon). This is noteworthy, indicating that the vertical distribution of dinoflagellates at these oceanic, tropical localities may not be influenced essentially by diurnal migration.

At Stations 1 and 2 the coccolithophorids decrease greatly between 50 and 100 m, while at Station 88 the greatest decrease in cell numbers is from 95 to 119 m, as with dinoflagellates.

Within the upper 50 m, and at Station 88 100 m, the coccolithophorids are uniformly distributed. This picture reflects the distribution of the predominant group within the coccolithophorids, as the vertical distributions of the species differ greatly. In Table 5 examples illustrating four types of vertical distributions of coccolithophorids

Table 5. The vertical distribution of some coccolithophorids

1. *Cyclococcolithus leptoporus*, *Coccolithus* sp., *Gephyrocapsa* sp.

Station 88		*Station* 1		*Station* 2	
Depth (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*
0	11,500	0	20,000	0	17,500
50	13,500	24	28,000	19	19,500
95	7500	52	21,000	43	6500 (?)
119	180	105	1712	87	220
159	20	158	384	130	920
217	20	211	250	175	224
323	328	318	348	267	265
434	192				

2. *Cyclococcolithus* cf. *fragilis*

Station 88		*Station* 1		*Station* 2	
Depth (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*
0	—	0	100	0	20
50	920	24	20	19	+
95	*1480*	52	400	43	—
119	780	*105*	*1180*	*87*	*720*
159	20	158	4	130	272
217	—	211	—	175	16
323	16	318	20	267	20
434	12				

3. *Thorosphaera flabellata*

Station 88		*Station* 1		*Station* 2	
Depth (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*
0	—	0	—	0	—
50	—	24	—	19	—
95	680	52	—	43	20
119	2060	105	8 (?)	87	80
159	*4340*	*158*	*708*	*130*	*256*
217	—	211	—	175	8
323	—	318	—	267	—
434	4				

4. *Thoracosphaera heimii*

Station 88		*Station* 1		*Station* 2	
Depth (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*
0	—	0	—	0	—
50	20	24	—	19	—
95	+	52	—	43	—
119	+	105	136	87	20
159	—	158	16	130	72
217	—	211	35	175	28
323	40	318	44	267	30
434	44				

are presented. The most numerous group, consisting of at least three small and very similar species, shows a rather even distribution down to 50 m at Stations 1 and 2 and to 95 m at Station 88. These species are restricted to the surface layer, with their maxima probably at a depth of 25 to 50 m. *Cyclococcolithus fragilis* had a rather pronounced maximum at about 100 m, with comparatively few cells above or below this level. *Thorosphaera flabellata* had its maximum still deeper, at about 150 m, with almost no cells below this depth. Since no cells occurred in the surface layers, it seems as if this species is confined to the layer between 100 and 150 m. The last distributional type as represented by *Thoracosphaera heimii* is probably restricted to below 100 m.

Coccolithophorids were recorded at rather great depths in the Mediterranean by LOHMANN (1902) and BERNARD (e.g. 1937), the latter finding a similar abundance of " small coccolithophorids " at 350 m as well as at 10 and 50 m.

The large coccolithophorid material from the *Deutschland*-Expedition showed that the different species of coccolithophorids had a different vertical distribution (LOHMANN, 1920). Some of them, such as *Deutschlandia anthos*, having a habitus similar to *Thorosphaera flabellata*, are characterized as " Dämmerungsformen."

In the present investigation *Thoracosphaera heimii* may be characterized as a " Dämmerungsform." This species was found by LOHMANN (l.c.) down to 200 m, but maximum abundance occurred at 50 to 100 m. Such discrepancies may depend upon the difference in the number of samples, and also demonstrate how careful one must be in drawing general conclusions from so few stations as in the present investigation. Another species belonging to the shadow flora (CHUN, 1900), *Halosphaera viridis*, was found only in very small numbers. There was an indication that deeper layers were preferred. The " Olivgrüne Zellen " recorded from the Atlantic (HENTSCHEL, 1928) were restricted to the deepest layers. They were also found in the Adriatic by SCHILLER (1931) who characterized them as " grüngelbe." SCHILLER's and HENTSCHEL's observations are based on unpreserved, centrifuged samples. In

Table 6. *The vertical distribution of* " *Olivgrüne Zellen* " (HENTSCHEL)

Station 88		*Station* 1		*Station* 2	
Depth (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*	*Depth* (m)	*Cells per litre*
0	—	0	—	0	—
50	—	24	20	19	—
95	—	52	60	43	—
119	—	105	—	87	20
159	—	158	232	130	296
217	—	211	630	175	1160
323	1600	318	1344	267	295
434	1296				

the formalin preserved samples from the equatorial Pacific they had a brown tint, but they were without doubt cells of the same kind as described by HENTSCHEL and SCHILLER. In the present study they were found in greater amounts than in the two other investigations, probably because of the difference in the concentration method. Table 6 gives the vertical distribution of these cells at the three stations. The cells were found mainly at depths below 150 m, and at Station 88 1600 cells per litre occurred

at 323 m. Unfortunately there was no possibility of supplementing HENTSCHEL's and SCHILLER's descriptions of these cells. It was only proved that they gave no nucleus reaction with acetic-carmine, this being in agreement with the observations of HENTSCHEL and SCHILLER.

The stability conditions are illustrated in Fig. 3. The only feature of importance in this discussion is the location of the thermocline. Only at Station 88 does the lower boundary of dinoflagellate and the coccolithophorid abundance coincide with the thermocline, situated at 97·5 m (determined by bathythermograph). At the two other stations the thermocline is probably too deep to influence vertical distribution of the phytoplankton. (At Station 1 the thermocline was recorded at 125 m and at Station 2 at 131 m). The shallower thermocline at Station 88 may contribute to the higher phosphate content in the surface layer at this station (see p. 40) as found by CROMWELL (1953) for the northern boundary of the counter current.

The observations presented from this tropical locality show that even in the tropics the bulk of the phytoplankton may be restricted to the upper 50-100 m, and that the number of cells below 150 m may be without importance when computing the standing stock. This corresponds well with the observations of STEEMANN NIELSEN (1957) of the depth of the photosynthetic layer (the depth where 1 per cent of the surface light is found) in the equatorial Pacific in March at 78-81 m. RILEY's (1957) assumption that sampling depths greater than 150 m might be needed for a compution of the standing stock would not seem to apply to the present area. The abundant species were found throughout the sampling column, but the specimens at greater depths, especially the diatoms, were present only as empty frustules. An exception was two cells of *Chaetoceros atlanticus* f. *neapolitanus* from 318 m which had normal chromatophores. The dinoflagellates, more numerous than the diatoms below about 150 m, were mostly full of cytoplasmic material at the greater depths, very often containing droplets probably consisting of storage-products.

The composition of the phytoplankton population

In textbooks dinoflagellates and coccolithophorids are considered to be the predominant components of the phytoplankton of the warmer zones, at least as far as number of species is concerned. The predominance of dinoflagellate species was convincingly demonstrated by RAMPI (1952), who from 3°S to 8°N and 164°W to 175°W found about 170 species, varieties or forms of dinoflagellates and about 40 species of diatoms. Coccolithophorids were not studied as the observations were based on material obtained by Clarke-Bumpus sampler. In the present material the coccolithophorids predominated in cell numbers in all samples except one. The diatoms were more abundant than the dinoflagellates except in those samples collected deeper than 150 m.

The numbers of species recorded for the three groups are given in Table 7, to which some unidentified species must be added to *Pennales*, *Dinokontae* (especially to the 'gymnodinians') and *Coccolithophoridaceae*. In this material the number of *Dinophyceae* exceeds that of the *Bacillariophyceae* although the difference is very small. The disagreement with RAMPI's results may be caused by the different sampling and counting techniques. In poor waters the net filters a large amount of water and thus may catch even the rarer species but in the present study only volumes of 50 or 250 ml were examined. (On the other hand, the loss of the smaller forms of

all groups of phytoplankton is especially pronounced in net-hauls from the plant-poor tropical waters, where the organisms are not abundant enough to produce a clogging effect and so retain the small forms.) For this reason the present study demonstrated the presence of so many coccolithophorids in the Pacific Ocean, as well as small dinoflagellate species, not recorded from this ocean previously. Another result shown in Table 7, also in disagreement with the commonly accepted view, is the

Table 7. Number of species

Bacillariophyceae (Diatoms)	*42*
Centrales (Centric forms)	24
Pennales (Pennate forms)	18
Dinophyceae (Dinoflagellates)	*53*
Desmokontae	18
Dinokontae	35
Coccolithophoridaceae (Coccolithophorids)	*34*

similarity in the number of centric and pennate diatom species. Usually in an oceanic plankton diatom population the centric forms will predominate in number of species, but in the present material the proportion between the number of species of centric and pennate forms is similar to that found in inshore, temperate localities.

Furthermore, all the centric diatom species appeared in rather small numbers, while up to 60,000 cells per litre of the pennate form, *Nitzschia delicatissima*, possibly mixed with another *Nitzschia* species, were found in a surface sample. These cells were usually clumped together and the number given is therefore approximate. Another pennate diatom, an unclassified *Thalassiothrix* (?), which was a constant component of the flora, also reached numbers only exceeded by those of the numerous 'gymnodinians' and coccolithophorids. The same was also recorded for still another unclassified pennate form (*Thalassionema nitzschioides*, very poorly silicified ?).

The only species or groups which may be ranked as numerous are *Nitzschia delicatissima*, 'small coccolithophorids' (including small *Cyclococcolithus leptoporus*; *Coccolithus* sp. and *Gephyrocapsa* sp.) and 'gymnodinians,' (listed according to relative abundance and with populations exceeding 10,000 cells per litre). *Nitzschia delicatissima* and 'gymnodinians' are also particularly abundant and predominate in inshore and temperate regions. *Nitzschia delicatissima* is probably ubiquitous, while the group 'gymnodinians' consists of species characteristic for each habitat.

Only one of the armoured dinoflagellates recorded, *Oxytoxum variabile*, about 20 μ long, was found in numbers greater than 1000 cells per litre (the highest recorded numbers were 8 in a 2 ml aliquot, 20 in a 50 ml aliquot, 39 in a concentrate of 250 ml).

Even by using a method that was suitable for detailing the smaller forms, no numerous dinoflagellate was observed. Small, numerous, armoured dinoflagellates, usually present in inshore waters, are probably restricted to such habitats, and the dinoflagellate population of a tropic, oceanic locality consists of many species, all represented by few specimens.

Species recorded from the Subantarctic and the equatorial region

The primary purpose of the present study was to compare the phytoplankton of the Subantarctic with that of a locality farther north in the Pacific Ocean. The following species were found in apparently similar state in the two regions,

Vol 6 1959

Diatoms

Chaetoceros aequatorialis
— atlanticus f. neapolitanus
— peruvianus
— tetrastichon
Dactyliosolen mediterraneus
Nitzschia bicapitata
— kolaizeckii
— sicula
Thalassiosira subtilis

Silicoflagellates

Dictyocha fibula var. messanensis

Dinoflagellates

Ceratium fusus var. seta
Exuviaella baltica
Goniaulax spinifera
— turbynei
Oxytoxum variabile

Coccolithophorids

Acanthoica quattrospina
Coccolithus huxleyi
Cyclococcolithus leptoporus

The diatoms *Rhizosolenia alata*, *Rh. cylindrus*, *Thalassionema nitzschioides* and *Thalassiosira antarctica* were also found in both regions, but were present in the equatorial region in very small numbers. There were only almost empty cells of *Rh. cylindrus* and *Th. antarctica*, two and one cells recorded respectively, found at relatively great depths (105 and 211 m). It is therefore assumed that the cells of these two species may have been transported from a distant area. *Rhizosolenia alata* and *Thalassionema nitzschioides* were probably represented by unlike varieties or forms in the two regions.

One or two of these species recorded as common to the two regions are ubiquitous, the other probably cosmopolitan warm water species.

It must be stressed that most of the common species, especially *Chaetoceros atlanticus* f. *neapolitanus*, *Ch. peruvianus* and *Thalassionema nitzschioides* were far more numerous in the Subantarctic than in the equatorial region. *Coccolithus huxleyi* was far more abundant in the Subantarctic than in the equatorial region. On the other hand, *Cyclococcolithus leptoporus* was common in the equatorial region but recorded in small numbers at a few of the stations in the Subantarctic.

In Table 8 all the species are listed which were identified during the counting or by examination under an ordinary microscope. Some of the species could easily be recognized but not properly identified. These are conferred to a given species or simply identified to genus only. A species given in brackets means that it was found only as empty frustules.

Further details about the geographical distribution of some of the recorded species will be published in another report, (HASLE, in preparation) where the new species in Table 8 will be described.

The existence or non-existence of seasonal variation

No information about seasonal variation of tropical phytoplankton is given in the literature either because of insufficient observations or because of negative results of such studies. MARSHALL (1933) found no seasonal variation in the phytoplankton from the Great Barrier Reef, while MOTODA (1941) found that his observations did not allow him to draw conclusions about the existence of seasonal variation in the phytoplankton. JOHNSON (1954) studied the phytoplankton of the northern Marshall Islands for part of the year and found some variations in the quantities.

Table 8 (a)

BACILLARIOPHYCEAE (Diatoms)

Centrales (Centric forms)
Actinocyclus ehrenbergii var.
— sp.
Asteromphalus flabellatus (Breb.) Grev.
Bacteriastrum elegans Pav.
Brenneckella lorenzenii Lohm.
Chaetoceros aequatorialis Cl.
— *atlanticus* var. *neapolitanus* (Schröd.) Hust.
— *peruvianus* Brightw.
— cf. *socialis* Laud.
— *tetrastichon* Cl.
Coscinodiscus crenulatus Grun.
— *excentricus* Ehrbg.
Coscinosira oestrupii Ostf.
Dactyliosolen mediterraneus Perag.
Planktoniella sol (Wall.) Schütt
Rhizosolenia alata Brightw.
— *bergonii* Perag.
— *cylindrus* Cl.
— *styliformis* Brightw.
Thalassiosira antarctia Comber
— (*decipiens*) (Grun.) Jørg.
— sp.
Triceratium cf. *formosum* Brightw.
— sp.
Pennales (Pennate forms)
Asterionella sp.
Nitzschia bicapitata Cl.
— cf. *closterium* (Ehrbg.) W.Sm.
— *delicatissima* Cl.
— *kolaizeckii* Grun.
— *pacifica* Cupp
— *sicula* (Castr.) Hust.
— sp.
Pseudoeunotia doliolus (Wall.) Grun.
Thalassionema elegans Hust.
— *nitzschioides* var. *inflata* Heid.
— — — *parva* Heid.
Thalassiothrix cf. *delicatula* Cupp
— cf. *frauenfeldii* Grun.
— *vonhoeffenii* Heid.
— sp.
Tropidoneis sp. 1
— sp. 2

Table 8 (b)

DINOPHYCEAE (Dinoflagellates)

Desmokontae
Amphisolenia globifera Stein
— *lemmermannii* Kof.
Dinophysis parva Schill.
— *umbosa* Schill.
Exuviaella aequatorialis n. sp.
— *baltica* Lohm.
— *compressa* Ostf.
— cf. *oblonga* Schill.
— cf. *pyriformis* Schill.
— *vaginula* (Stein) Schütt
Phalacroma mucronatum Kof. et Skogsb.
— *ovum* Schütt
— *ruudii* Braarud
Porella sp.
Prorocentrum maximum Schill.
— *obtusidens* Schill.
— *rostratum* Stein
Triposolenia bicornis Kof.
Dinokontae
Blepharocysta cf. *paulsenii* Schill.
Ceratium candelabrum f. *commune* Böhm
— *declinatum* f. *debile* Jørg.
— *fusus* var. *seta* (Ehrbg.) Jørg.
— *trichoceros* (Ehrbg.) Kof.
Epiperidinium michaelsarsii Gaarder
cf. *Glenodinium danicum* Pauls.
Goniaulax braarudii n. sp.
— *minima* Matz.
— *polygramma* Stein
— *spinifera* (Clap. et Lach.) Dies.
— *turbynei* Murr. et Whitt.
Gymnodinium aequatoriale n. sp.
— *oceanicum* n. sp.
Murrayella spinosa Kof.
Oxytoxum caudatum Schill.
— *curvatum* (Kof.) Kof.
— *elegans* (?) Pav.
— *laticeps* Schill.
— *milnerii* Murr. et Whitt.
— *sceptrum* (Stein) Schröd.
— *scolopax* Stein
— *variabile* Schill.
Peridinium brevipes Pauls.
— *globulus* var. *quarnerense* Br. Schröd.
— *minusculum* Pav.
— *tenuissimum* Kof.
Podolampas palmipes Stein
— *spinifer* Okamura
cf. *Pronoctiluca acuta* (Lohm.) Schill.
Protoceratium areolatum Kof.
— *reticulatum* (Clap. et Lach.) Bütschli
cf. *Protoerythropsis* sp.
cf. *Pyrocystis* sp.
cf. *Pyrophacus* sp.

Vol 6 1959-

Table 8 (*c*)

COCCOLITHOPHORIDACEAE
(Coccolithophorids)

Acanthoica quattrospina Lohm.
Anoplosolenia sp.
Anthosphaera sp.
Calciosolenia sp.
Calciopappus (?) sp.
Calyptrosphaera cf. *quadridentata* Schill.
Coccolithus huxleyi (Lohm.) Kpt.
Cyclococcolithus fragilis (Lohm.) Gaarder
— *leptoporus* (Murr. et Blackm.) Kpt.
— *mirabilis* (Lohm.) Kpt.
— *sibogae* (Weber van Bosse) Gaarder
Deutschlandia sp.
Gephyrocapsa sp.
Halopappus sp.
Helicosphaera carterii (Wall.) Kpt.
Lohmannosphaera sp.
Michaelsarsia sp.
Ophiaster sp.
Pontosphaera discopora (?) Schill.
— *nana* Kpt.
— *sessilis* Lohm.
Rhabdosphaera erinaceus Kpt.
— *stylifer* Lohm.
— cf. *tignifer* Schill.
Syracosphaera mediterranea Lohm.
— *molischii* Schill.
— *pulchra* Lohm.
— *tuberculata* Kpt.
— sp. 1
— sp. 2
Thoracosphaera heimii (Lohm.) Kpt.
Thorosphaera flabellata Halldal et Markali
Umbellosphaera irregularis Paasche

Table 8 (*d*)

OTHER FLAGELLATES etc.

Chilomonas marina (Braarud) Halldal
Danasphaera indica Steemann Nielsen
Dictyocha fibula var. *messanensis* (Haeckel) Lemm.
Halosphaera viridis Schmitz
Monosiga marina Grøntved
" Olivgrüne Zellen " Hentschel
cf. *Peranema* sp.
Pterosperma cristatum Schill.
— *parallelum* Gaarder
Solenicola setigera Pav.

CILIATES
Didinium parvulum Gaarder
Acanthostomella sp.
Cantharielia cf. *brevis* Kof. et Campb.
Cyttarocylis longa Kof. et Campb.
Dadayiella ganymedes (Entz, Sr.) Kof. et Campb.
Dictyocysta sp.
Eutintinnus spp.
Ormosella bresslaui Kof. et Campb.
Parundella cf. *inflata* Kof. et Campb.
Poroecus curtus Kof. et Campb.
Proplectella ellipsoida Kof. et Campb.
— cf. *perpusilla* Kof. et Campb.
Protorhabdonella simplex (Cl.) Jørg.
Salpingella cf. *costata* (Laack.) Kof. et Campb.
Steenstrupiella gracilis (Jørg.) Kof. et Campb.
— *steenstrupii* (Clap. et Lach.) Kof. et Campb.
Strombidium acuminatum (Leeg.) Kahl
— *emergens* (Leeg.) Kahl
— *obliquum* Kahl
— *strobilus* (Lohm.) Kahl

None of these three investigations represent an oceanic locality, and the variations found in the last of them may reflect only local differences resulting from water movements. In his two-year study of the Sargasso Sea (viz. the subtropics) RILEY (1957) was able to demonstrate a distinct quantitative seasonal variation in the phytoplankton.

Considering the equatorial Pacific, one has to resort to available observations of the ecological factors and indirect evidence from Secchi disk readings. The amount of radiation reaching the sea surface in the equatorial region is more constant than in other parts of the world. The difference between the greatest and least amount is less than half of what it is at a latitude of 52°S (after Kimball's data from SVERDRUP *et al.*, 1946 p. 103). The sea temperature at the equator is also constant. The amplitude is less than 2°C (DIETRICH and KALLE, 1957 p. 142). As in most other oceanic localities the salinity is constant throughout the year. Phosphate measurements in this region have been made for a number of years, but the data have not yet been studied with a

view to looking for seasonal variations. Such a study will be complicated by the presence of divergences and convergences near the equator and their variations in strength and position.

The apparent seasonal difference in zooplankton abundance which was found by KING and DEMOND (1953) is in contrast to these negative findings. They assumed that in these waters there was no chance of a seasonal variation in the phytoplankton. But seasonal difference in zooplankton abundance might give rise through grazing to a parallel variation in the phytoplankton.

In tropical, oceanic waters mostly free of terrigenous particles, Secchi disk readings may be expected to give a good indication of the abundance of phytoplankton. From the central equatorial Pacific there exist Secchi disk readings from October-December 1955 (KING *et al.* 1957) and August-September 1956 (AUSTIN, 1957). The results are :

Date	Latitude	Longitude	Secchi disk depth	Depth for 1 % of surface light	Number of observations
29 Sept.–12 Dec. 1955	8°S–12°N	111°W–162°W	18 m–33 m	49 m–88 m	49
17 Aug.–14 Sept. 1956	13°S–0°	131°W–140°W	16 m–28 m	43 m–77 m	25

The depth for 1 per cent of surface light is calculated from the extinction coefficient $k = 1{\cdot}7/D$ (D is the Secchi disk depth) (POOLE and ATKINS, 1929) and $L = 1/k \times 4{\cdot}6$ (the last expression computed from $I/I_0 = e^{-kL}$, where I_0 = initial intensity, I = intensity at depth, L, in metres).

The minimum values indicate a rather abundant population, at least compared with data from the subantarctic Pacific (Table 4). This may mean that March is a time with scarcer phytoplankton than August–December. This assumption is supported by the measurement of the depth of the photosynthetic layer made by the *Galathea*-Expedition from 13°S to 13°N in the central Pacific. These were made in March and gave depths from 78 to 120 m (STEEMANN NIELSEN, 1957 Table 20), that is, depths greater than those found from August to December.

The effect of the equatorial upwelling

A positive correlation between the upwelling in the vicinity of the equator, nutrient enrichment and zooplankton abundance was demonstrated by KING and DEMOND (1953) and HIDA and KING (1955). The greatest abundance of tuna fish larvae was found somewhat farther north (SETTE, 1955).

If the Secchi disk depths are examined more thoroughly, especially those of KING *et al.* (1957), a band with higher extinction is distinguished near the equator. But it is also evident that this band varies in extent and position along the section from 110°W to 160°W, and that in between these low values of the Secchi disk depths the highest ones are also found. The four measurements made by the *Galathea*-Expedition from about 3°S to 13°N also indicate higher extinction in this area than at higher latitudes (depth of the photosynthetic layer, 78 m–85 m). Compared with Secchi disk depths from other tropical waters as summarized by CLARKE (1954) and DAS (1954), those of the central equatorial Pacific are low. BIRT (1958) measured light-transmission

Vol 6 1959

in tropical Pacific waters and found that the equatorial region had the lowest but one transmission of five areas sampled. The area south-west of the Galapagos Islands, the Equatorial Countercurrent and the North Equatorial Current had higher transmissions ; only the Peru Current had lower.

The Swedish Deep-Sea Expedition found a great amount of particulate matter near the equator, especially between about 2°S to 2°N in the cross-section at 135°W (JERLOV, 1953). As the supply of terrigenous components must be low in this area, the particles were interpreted largely as representing phytoplankton populations and plankton remnants. The effect of upwelling on the productivity was demonstrated by DOTY *et al.* (1956) who found high productivity in the waters in the vicinity of the equator. The observations by the *Galathea*-Expedition in this area did not demonstrate a direct correlation between rate of production and phosphate content, which is perhaps a reflexion of the paucity of observations (STEEMANN NIELSEN, 1957).

In view of the variation in zooplankton and depth of the photosynthetic layer throughout the year as discussed above, and the productive fertile zone near the equator, a quantitative phytoplankton investigation, using the same sampling and counting methods as in the present study, but far more extensive in time and space, should yield valuable results.

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Vol 6 1959

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Vol.
6
959-60

SHORTER COMMUNICATION

The temperature correction of oceanic inorganic phosphorus analyses

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Abstract—The colour intensity of the molybdate-blue complex in the analysis for inorganic phosphorus in sea-water increases 1·25 per cent per degree centigrade. The apparent phosphorus content of sea-water samples analysed in tropical regions may be erroneously high unless temperature is controlled or corrections are made. The effect of temperature correction on the agreement between total and inorganic phosphorus contents of Atlantic Ocean water is described. It is recommended that temperature corrections be included in future publications of data on the distributions of inorganic phosphorus.

Vol
6
1959

INTRODUCTION

THE intensity of colour developed in the Deniges-Atkins method (1923) for the determination of phosphorus increases with temperature. ROBINSON and THOMPSON (1948) and WOOSTER and RAKESTRAW (1951) found increases in colour of about 1 per cent per degree C. HARVEY (1955) cites a value of 1·2 per cent per °C for the colour increment. In oceanographic studies of the distribution of phosphorus, considerable variation in the correct value may be expected if the temperature variation among the samples analysed is great. For this reason it has become customary to allow the samples to reach the ambient temperature before analysis, thus eliminating the effects of the large temperature variations between surface and deep water, which may be as much as 25–30°C in tropical and subtropical areas. WOOSTER and RAKESTRAW (1951) recommend a constant temperature water bath for this purpose. The temperature effect would also be eliminated if the standard and unknown solutions are at the same temperature during comparison (for visual estimation of colour) or calibration (for photometric determinations).

The temperature effect has commonly been neglected in oceanographic work, presumably because it has been assumed to be a small error in comparison with other inaccuracies of the method. During recent extensive determinations of the distribution of phosphorus in the Atlantic Ocean in connexion with observations made during the International Geophysical Year, we have found it essential to make temperature corrections for the inorganic phosphorus analyses made on shipboard. Correction was necessary because the calibration of the photometer, though repeated frequently, was performed when temperatures were about 20°C, whereas the analyses were performed at temperatures ranging up to 33°C in the tropics. The validity of the temperature correction is indicated by comparisons between the inorganic and total phosphorus.

METHODS

Inorganic phosphate was determined by the Deniges–Atkins method as described by WOOSTER and RAKESTRAW (1951). The colour developed by the molybdate-blue complex was measured in the photo-electric colorimeter (FORD, 1950) using a red light filter (Corning 2408) and a cell with a 29 cm light path for concentrations less than 1·0 μg A/l or a cell with a 9 cm light path for higher concentrations. Samples were analysed after they had been adjusted to the laboratory temperature on shipboard. The photometer was calibrated using standard solutions prepared with surface Sargasso sea-water. This water was nearly phosphate-free and the small residual phosphorus content was determined on blank samples.

Total phosphorus was determined by the HARVEY method as modified by KETCHUM *et al.* (1955), using a tube length of 29 cm in the photometer. The calibration was made by subjecting standards, prepared in Sargasso Sea-water, to the same procedure used for the unknowns.

OBSERVATIONS

The need for temperature correction first became apparent when the inorganic and total phosphorus concentrations in the deep water of the Atlantic Ocean were compared. The data were collected on Cruise 10 of the Research Vessel *Crawford* as part of our observations for the programme of the International Geophysical Year. In such comparisons, particulate or dissolved organic phosphorus

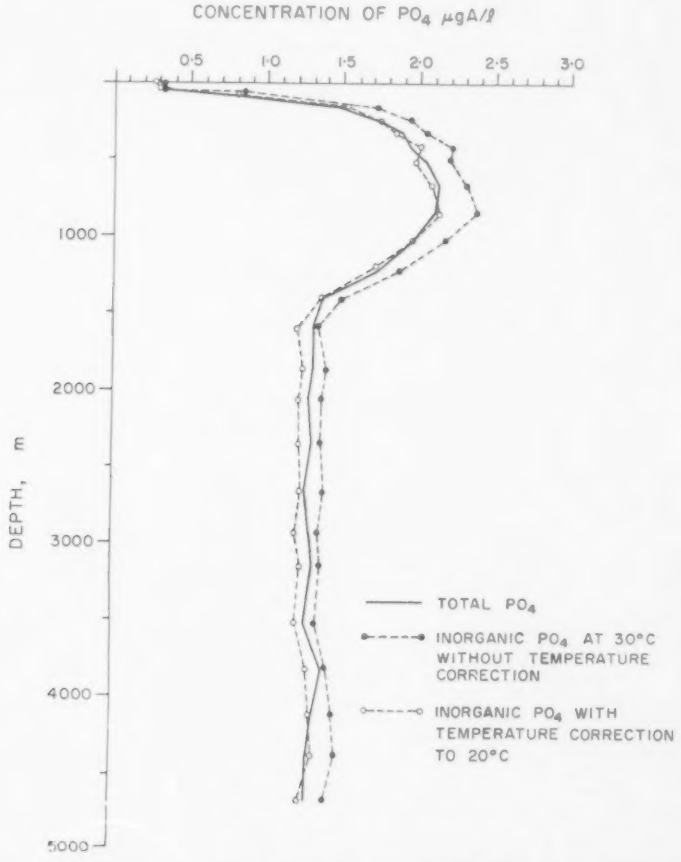

Fig. 1. Comparison of the apparent inorganic with the total phosphorus for a station in the tropical Atlantic. (*Crawford* Cruise 10, Station 168, Lat. 8°N, Long. 36°W).

is presumed to be present in the water when the total concentration significantly exceeds that of inorganic phosphorus. Consistently, however, in the deep water of tropical and sub-tropical regions, the inorganic concentration exceeded that of the total, a result which is theoretically impossible. It was apparent, therefore, that there was a consistent bias in our analytical results. Since the discrepancy was exaggerated in the tropics, where shipboard laboratory temperatures averaged about 30°C, the effect of correction for temperatures has been studied.

The analyses for inorganic phosphorus were made on shipboard as soon as possible after collection, but after allowing ample time for the samples to come to laboratory temperature. The temperature

of each sample was not recorded on this cruise, but periodically the ambient laboratory temperature was recorded. The excess in temperature over the calibration temperature (20°C) was used to correct the inorganic phosphorus concentrations. The correction applied was 1 per cent per °C in accordance with data of ROBINSON and THOMPSON (1948) and WOOSTER and RAKESTRAW (1951). Since the total phosphorus was determined in Woods Hole at the calibration temperature no correction for these was made.

An example of the effect of this correction on the apparent phosphorus at various depths is given in Fig. 1. These data are for *Crawford* Station 168 at 8°N, 36°W in the central Atlantic. The phosphorus analyses on shipboard were made at 30°C, and the corrected values are thus decreased by 10 per cent. The uncorrected inorganic phosphorus exceeded the total at all depths. Although the difference on individual samples is not statistically significant (KETCHUM *et al.*, 1955), the consistency of the bias is apparent. Applying the temperature correction produced complete agreement in the layer of maximum concentration. In the deep waters this correction reduced the inorganic concentrations to values slightly less than the total.

As a further illustration of the effect of the temperature correction, the average inorganic and total phosphorus concentrations for samples collected at 2000 (± 150) m on four sections are compared in Table 1. This depth was selected to conform to REDFIELD's (1958) presentation of the global

Table 1. *Comparison of total and inorganic phosphorus in sea-water at* 2000 m *depth in the Atlantic Ocean*

Section	*Number of Samples*	*Inorganic* $PO_4^{=}$ (μg A/1) *Uncorrected* Mean	σ	*Corrected* Mean	σ	*Total* P (μg A/1) Mean	σ
Western Atlantic Basin							
Bermuda–Dakar	6	1·35	0·080	1·27	0·068	1·27	0·057
8° 15′N	8	1·42	0·100	1·22	0·091	1·23	0·074
8° 15′S	11	1·55	0·040	1·36	0·036	1·36	0·052
15° 45′S	18	1·60	0·094	1·47	0·091	1·49	0·093
Eastern Atlantic Basin							
Bermuda–Dakar	6	1·46	0·088	1·37	0·088	1·39	0·069
8° 15′N	13	1·47	0·104	1·34	0·078	1·35	0·070
8° 15′S	15	1·53	0·117	1·36	0·035	1·42	0·088
15° 45′S	9	1·62	0·141	1·49	0·100	1·53	0·072

Table 2. *Effects of temperature on the development of colour in the phosphorus analysis* (9cm *tube for inorganic and* 29 cm *tube for total analysis*)

°C	*No. of samples*	*Average factor* μg A/1	*Standard deviation* σ
Inorganic analysis			
15·0°	10	$1{\cdot}228 \times 10^{-2}$	$0{\cdot}056 \times 10^{-2}$
23·5°	9	$1{\cdot}110 \times 10^{-2}$	$0{\cdot}036 \times 10^{-2}$
30·0°	9	$1{\cdot}016 \times 10^{-2}$	$0{\cdot}029 \times 10^{-2}$
Total Analysis			
15·0°	10	$0{\cdot}874 \times 10^{-2}$	$0{\cdot}058 \times 10^{-2}$
23·5°	8	$0{\cdot}820 \times 10^{-2}$	$0{\cdot}090 \times 10^{-2}$
30·0°	10	$0{\cdot}793 \times 10^{-2}$	$0{\cdot}070 \times 10^{-2}$

distribution of phosphorus. The standard deviations for these data include not only analytical variations but actual differences in geographical distribution. The uncorrected inorganic phosphorus concentrations exceeded the total in all cases, emphasizing again the consistency of the bias. The corrected inorganic phosphorus concentrations closely agree with the total in every section.

The effect of temperature on colour development has been redetermined using our photometer and methods. Complete calibrations were performed by adding inorganic phosphorus in concentra-

Vol 6 1959

tions of 0·5, 1·0, 1·5, 2·0 and 2·5 μg atoms per litre to Sargasso Sea-water. These samples were analysed after equilibration at temperatures of 15°, 23·5° and 30°C in constant temperature water baths. For each sample, a factor which relates each unit of the arbitrary scale of the photometer with microgramme atoms per litre was computed. The results are given in Table 2 and in Fig. 2.

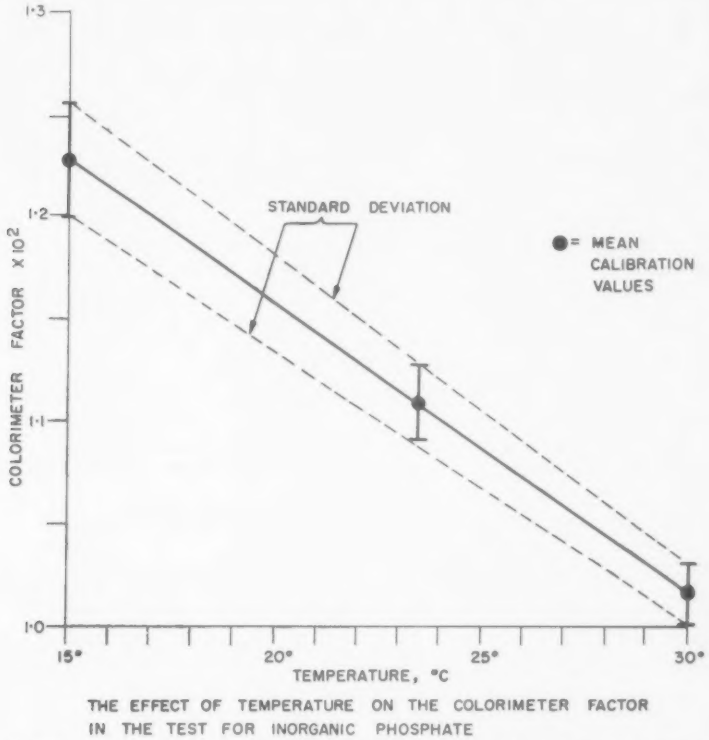

Fig. 2. Calibration for inorganic phosphorus analysis as a function of temperature. The dotted lines show the standard deviation.

The increased colour development, which produces a lower factor, corresponds to 1·25 per cent per degree centigrade. A similar set of calibrations was performed on the total phosphorus analysis at these same temperatures. The temperature correction in this method corresponds to 0·65 per cent per °C. As mentioned above, it has not been necessary to correct the total values because both calibration and analysis are performed at substantially the same temperature.

Table 3. *Apparent phosphorus content of a sea-water sample to which* 2·0 μg A/l *of inorganic phosphorus was added. (For explanation see text)*

Temperature of analysis	*Uncorrected* μg A/l	*Corrected to* 20°C μg A/l	*Observed* μg A/l
15·0°	1·94	2·04	2·08
23·5°	2·14	2·07	2·04
30·0°	2·31	2·07	2·05

The validity of the temperature correction is further indicated by the data in Table 3 which compares the apparent phosphorus concentration of a sample analysed at these different temperatures and computed in three ways. The uncorrected values are obtained by computing the phosphorus content

using the calibration factor measured at 20°C. This corresponds to the uncorrected values in Fig. 1. In the second column these values have been corrected to 20°C by application of the 1 per cent/°C correction, as was done with the *Crawford* data. In the final column the concentration is computed using the average calibration factor derived for the appropriate temperature. Both methods of correcting for temperature of analysis appear to be satisfactory.

CONCLUSIONS

To obtain correct values for inorganic phosphate measured at sea a temperature correction must be applied unless the colour development in standards and unknowns is measured at the same temperature. This was, of course, true when direct comparisons were made visually. The error may be considerable, however, when a photometer is used.

Several methods are available to eliminate this error. Samples can be placed in a constant temperature bath and brought to the calibration temperature, as suggested by WOOSTER and RAKESTRAW (1951). In the humid tropics, however, condensation on the walls of the absorption cell can give trouble unless the temperature is higher than is expected at any time during the cruise. Furthermore, on small ships and long cruises, available space may not permit this refinement. Another possibility is to include standards and to determine the calibration factor separately for each set of analyses. When large numbers of samples are to be analysed by a single chemist on the ship this is also impractical.

The simplest method for correcting for the effect of temperature involves calibrating the photometer at various temperatures, and using the appropriate calibration factor for the temperature of each sample at the time of analysis. From the calibration data in Table 2 and Fig. 2, a table of calibration factors at various temperatures, in 0·5°C intervals, was prepared for this purpose. The stability of the photometer and reagents can be checked during the cruise at any time regardless of temperature.

Temperature corrections have been applied to all of the inorganic phosphorus analyses collected by this Institution during the I.G.Y. For the earlier cruises, *Crawford* 10 and 16, a correction of 1 per cent/°C in excess of 20°C was applied. It was necessary to use the ambient air temperature in the laboratory for this purpose since on these cruises the temperature was not measured on each sample. For the later cruises, *Crawford* 22 and *Atlantis* 247, the calibration factor for the temperature of each sample has been used as described above. It is recommended that all inorganic phosphorus analyses made by photometer at sea be corrected in this way, or, if this is not done, that temperatures during calibration and analysis be recorded.

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Vol
6
1959

LETTERS TO THE EDITORS

Enrichment experiments as a means of studying nutrients limiting to phytoplankton production

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By inference from indirect evidence, it is generally agreed that nutrients limit photosynthesis in most of the oceans, particularly when the surface waters are thermally stratified. Phytoplankton growth is usually accompanied by a decrease in the concentrations of nitrate, phosphate, and silicate, and these elements sometimes reach levels during the summer which are undetectable by normal analytical procedures.

From the composition of the mineral nutrients in plankton algae relative to carbon, one may calculate from photosynthesis measurements the rate at which the nutrients must be supplied to the plants. This frequently reveals that their supply at any given time is adequate for no more than a few days growth. Unfortunately, we know little about the rate at which these materials are supplied to the surface waters by vertical mixing, and almost nothing of the rate at which they are regenerated through bacterial decomposition, animal excretion, etc. *in situ* in the surface waters. Thus conclusions concerning a dynamic process, growth, are often based entirely upon observations of a static property, the instantaneous nutrient concentrations.

Even here, however, the complete picture is usually lacking. Chemical nutrient studies accompanying biological investigations in the ocean usually include phosphate, frequently nitrate, occasionally silicate. There are a number of other elements essential to plant growth which are not major constituents of sea-water but which are seldom measured in relation to plant growth. Examples of these include iron, copper, manganese, cobalt, and the vitamins.

Furthermore, nitrate and phosphate measurements do not necessarily give the complete picture of the available nitrogen and phosphorus. In addition to nitrite and ammonia (which are sometimes measured and are usually at very low concentrations) both elements may be present as dissolved organic compounds of unknown identity in concentrations as high as or higher than the inorganic fractions. We can arrive at no conclusions regarding the biological significance of these compounds without knowing what they are. But it is not improbable that some of them may be used directly, if not preferentially, as a source of nutrients by some species of phytoplankton.

For these various reasons, then, the nutrient chemical observations normally made at sea rarely, if ever, provide sufficient information to draw qualitative or quantitative conclusions concerning the specific substances which limit phytoplankton production. We have therefore attempted to attack this problem by a more direct, experimental approach. These experiments entailed adding various nutrients to samples of surface water containing its natural phytoplankton and measuring their rate of uptake of $^{14}CO_3^=$ in an illuminated incubator (1500 ft candles) for 24 hours. As controls, we have employed unenriched surface water. In addition, since deep water invariably supports high levels of production when brought to the surface (i.e., in upwelling areas), we have used as a second control samples containing 1/2 surface water and 1/2 filtered deep water, doubling the observed ^{14}C uptake in these samples to make them comparable to the others which contained all surface water. The objectives in these experiments were twofold : (1) to see whether we could simulate the productive capacity of deep water by artificial enrichment and (2) to determine which of the added nutrients were critical in limiting ^{14}C uptake in the unenriched surface water.

The nutrients which were employed for these experiments are those which are used routinely in this laboratory for the culture of marine phytoplankton, but their concentrations have been adjusted to levels which for the most part are comparable to, though somewhat higher than, the maximum concentrations of these elements in sea-water. The complete medium consists of nitrate, phosphate,

silicate, thiosulphate, trace metals (including iron), and vitamins, (Table 1). The experiments consisted of the two controls described above, one sample enriched with the complete medium and six

Table 1. *Medium used in enrichment experiments*

1. $NaNO_3$	50 μgA N/litre
2. $Na_2PHO_3 \cdot 12H_2O$	5 μgA P/litre
3. $Na_2SiO_3 \cdot 9H_2O$	50 μgA Si/litre
4. $Na_2S_2O_3 \cdot 5H_2O$	8 μgA S/litre
5. Vitamin mix	
Thiamin . HCl	0·2 mg/litre
Biotin	1·0 μg/litre
B_{12}	1·0 μg/litre
6. Iron and trace metal mix	
Fe-sequestrine	1·30 mg/litre Fe
$CuSO_4 \cdot 5H_2O$	0·005 mg/litre Cu
$ZnSO_4 \cdot 7H_2O$	0·010 mg/litre Zn
$CoCl_2 \cdot 6H_2O$	0·005 mg/litre Co
$MnCl_2 \cdot 4H_2O$	0·100 mg/litre Mn
$Na_2MoO_4 \cdot 2H_2O$	0·005 mg/litre Mo

additional samples enriched with the same medium but each with one successive constituent omitted. At each station, then, there were 9 separate ^{14}C experiments which may be summarized as follows :

1. Surface water.
2. 1/2 surface water and 1/2 filtered deep water (results doubled).
3. Surface water and complete enrichment medium.
4. Surface water and enrichment medium with NO_3^- omitted.
5. Surface water and enrichment medium with $PO_4^{\equiv}$ omitted.
6. Surface water and enrichment medium with $SiO_3^{\equiv}$ omitted.
7. Surface water and enrichment medium with $S_2O_3^{\equiv}$ omitted.
8. Surface water and enrichment medium with Fe + trace metals omitted.
9. Surface water and enrichment medium with vitamins omitted.

The dark uptake of ^{14}C was not measured in these experiments, so the absolute rates of photosynthesis could not be calculated. The strength of the ^{14}C was approximately 10 μ curies. The samples enriched with iron and trace metals developed a slight precipitate which was removed on filtration. The counts are corrected for ^{14}C adsorbed on the filters and on this precipitate by means of appropriate blank experiments with filtered, sterile water.

This preliminary report will describe the results of enrichment experiments at 13 stations occupied on a section between Montauk Pt., New York and Bermuda on the oceanographic cruise *USNS Chain II* in December, 1958. The first seven stations included that part of the section across the continental shelf and into the Gulf Stream. The remaining six stations were located in the Sargasso Sea. The deep water used for the enrichment experiments was collected from 100 m or less at Stations 1-10, from 1000 m or more at Stations 11-16.

Table 2 shows the station numbers, positions, sonic depths, depth of the wind mixed layer (from bathythermograph records), depths from which the deep samples were taken for the enrichment experiments, and the concentrations of phosphate, nitrate plus nitrite, and silicate in the surface and deep samples. Table 3 shows the results of the individual enrichment experiments with photosynthesis (^{14}C uptake) reported as a percentage relative to the maximum carbon uptake observed in each experiment. This permits results to be compared from station to station, and to be averaged.

While the results from one station to the next were somewhat variable and inconsistent, certain trends in the data are obvious. Stations 1-10 were significantly different from Stations 11-16, so these two groups have been averaged separately. For the first group (shelf, slope, and Gulf Stream), the surface water controls averaged about 75 per cent of the samples enriched with complete medium. Adding subsurface water from depths to 100 m did not increase photosynthesis of the surface water. The omission of nitrate, phosphate, thiosulphate, and vitamins produced results not appreciably different from the complete enrichment, but the omission of silicate gave results which may be

Vol 6 1959

Vol. 6 959-60

Table 2. Description of the stations and water conditions where enrichment experiments were made (chemical concentrations in μgA/l)

Sta. No.	*Position*		*Depth (m)*			PO_4		$NO_3 + NO_2$		SiO_3	
			Bottom	*Wind mixed*	*'Deep sample'*	*Surface*	*'Deep'*	*Surface*	*'Deep'*	*Surface*	*'Deep'*
1	40° 57'N	71° 48'W	40	40	30	0·77	0·81	1·28	1·33	1·18	1·41
2	40° 43'N	71° 40'W	70	70	60	0·89	0·89	0·78	2·13	2·81	2·70
3	40° 31'N	71° 34'W	67	67	60	0·69	0·70	3·24	3·88	4·12	3·76
7	39° 36'N	70° 56'W	2248	25	100	0·51	0·92	2·19	13·2	2·35	6·23
8	38° 49'N	70° 19'W	2862	61	100	0·33	0·65	1·51	7·65	2·35	4·70
9	38° 00'N	69° 39'W	4018	76	100	0·22	0·55	0·87	4·67	2·35	2·59
10	37° 10'N	68° 30'W	4574	25	100	0·08	0·30	0·18	0·23	1·06	1·06
11	36° 25'N	68° 02'W	4850	153	4785	0·12	1·35	0·16	—	1·18	32·81
12	35° 36'N	67° 20'W	4978	67	1000	0·13	1·35	0·30	14·6	—	13·52
13	33° 11'N	65° 15'W	4744	110	1000	0·13	1·51	0·08	14·9	1·18	15·11
14	33° 03'N	64° 58'W	4830	76	1000	0·13	1·34	0·08	16·1	1·18	13·88
15	33° 58'N	65° 49'W	5075	—	1000	0·10	1·51	0·12	14·8	1·29	14·94
16	34° 48'N	66° 30'W	5133	159	1000	0·14	1·48	0·38	18·0	1·76	15·41

Table 3. Relative uptake of ^{14}C (24 hours at 1500 ft candles illumination) of enriched and unenriched surface water samples. (Headings of columns 5-10 represent substances omitted from the enrichment medium in each case)

Sta. No.	*Control*	*Deep water*	*Complete medium*	$-NO_3$	$-PO_4$	$-SiO_3$	$-S_2O_3$	*– Vitamins*	*– Fe + Trace metals*
1	79	60	100	76	91	87	92	92	77
2	68	75	89	100	87	76	96	89	64
3	67	77	82	100	91	95	85	96	62
7	100	84	99	99	96	83	98	99	—
8	84	70	94	100	99	95	89	91	63
9	30	39	93	91	91	57	100	91	91
10	49	67	87	90	100	87	70	80	43
Mean (1-10)*	72	71	98	100	99	87	95	96	71
11	30	100	95	89	98	59	86	100	39
12	50	94	83	100	87	60	83	77	43
13	8	100	58	57	80	12	57	100	32
14	16	79	100	84	80	89	25	100	39
15	12	69	100	76	46	37	96	90	14
16	26	81	85	92	100	40	67	88	51
Mean (11-16)*	26	95	94	90	89	53	74	100	39

*—to a maximum of 100 in each case.

Vol. 6 1959

significantly lower. However, the enrichment with everything but the iron-trace metal mixture had exactly the same effect as no enrichment at all.

In the Sargasso Sea samples, the unenriched controls assimilated only about one quarter as much ^{14}C as the samples enriched with complete medium. Again the omission of nitrate, phosphate, and vitamins did not subtract from the effectiveness of the enrichment mixture. Thiosulphate may have helped somewhat, while if silicate was left out, only about half as much photosynthesis occurred as in the complete medium. The omission of iron and trace metals again had the most pronounced effect, reducing photosynthesis almost to the level of the controls. In this group of experiments, the addition of water from 1000 m or more of depth stimulated photosynthesis to almost exactly the same extent as did the enrichment with the complete artificial medium.

The fact that enrichment with deep water did not stimulate photosynthesis at the first group of stations is not surprising, since the water in most cases was wind-mixed to approximately the same depths as that from which the samples were taken, and, therefore, the surface and ' deep ' water were essentially the same. This is borne out by the relative concentrations of nitrate, phosphate, and silicate at the two depths (see Table 2). The same criteria demonstrate the dissimilarity between the truly deep water (1000 m or more) and the surface water at Stations 11-16, and help explain the stimulatory effect of the former upon photosynthesis.

But it was the artificial enrichment experiments, particularly at the Sargasso Sea stations, that produced the most surprising results. Although the surface concentrations of nitrate and phosphate were extremely low, the addition of these elements apparently had no effect on photosynthesis. This recalls the earlier statement that the phytoplankton may be dependent upon the rate of regeneration of these elements and/or their presence as dissolved organic compounds, and more-or-less independent of their instantaneous concentrations as inorganic salts.

Rather than nitrogen and phosphorus, the nutrients limiting photosynthesis in these waters appear to be silicate and one or more components of the iron-trace metal mixture. This does not imply, of course, that if these substances were present in excess, nitrogen and phosphorus would not eventually, perhaps very quickly, become the limiting factors. It does imply, however, that they would become limiting at a higher rate of photosynthesis, which is another way of saying that the rate of primary production in the Sargasso Sea, at the time of our observations, was limited by nutrients other than nitrogen and phosphorus.

The purpose of this paper is to point out the possibilities of this approach in studying nutrient relations to photosynthesis in the sea. The results of our preliminary experiments, while suggestive, are not in themselves adequate to establish new ideas or refute conventional ones. The obvious next step is to separate and distinguish the critical elements in the iron-trace metal mixture. With this and perhaps other refinements to the technique, seasonal and regional studies of the specific nutrient elements limiting primary oceanic production may be undertaken.

Acknowledgements—The authors were assisted in the field by R. B. Williams. The chemical analyses reported in Table 2 were carried out by N. Corwin, R. F. Vaccaro and F. A. Richards.

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Observations of deep water overflow across the Iceland–Faroe Ridge

(*Received* 17 *February* 1959)

In recent years attention has been drawn to the problem of overflow into the Atlantic of the deep cold water in the Arctic basin (Cooper, 1955 ; Dietrich, 1956). This letter gives a brief description of an overflow across the Iceland-Faroe Ridge observed on the 3rd and 4th of August, 1958.

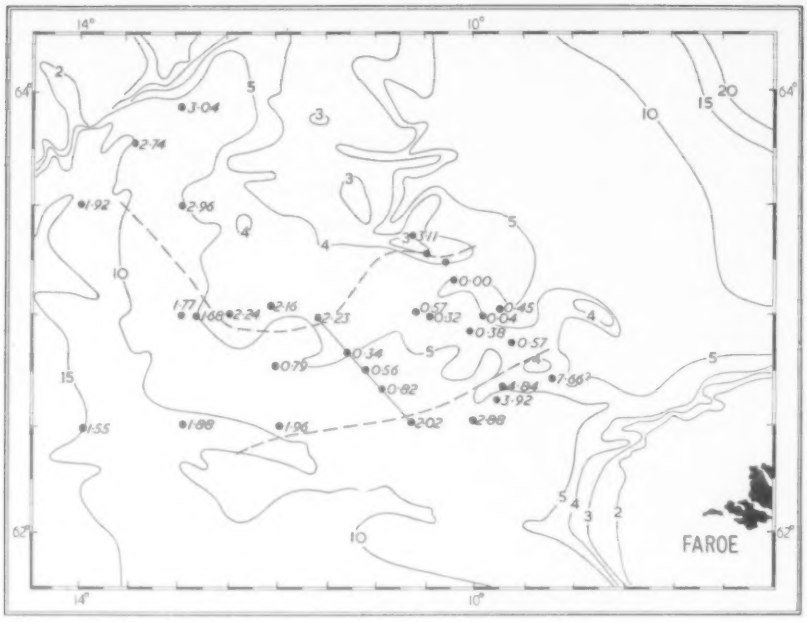

Fig. 1. Bottom temperatures on the Ridge. Aug. 3-4, 1958.

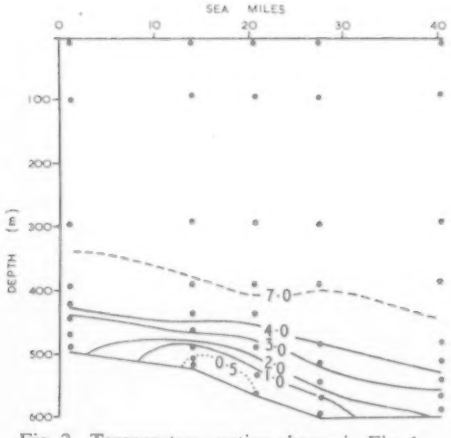

Fig. 2. Temperature section shown in Fig. 1.

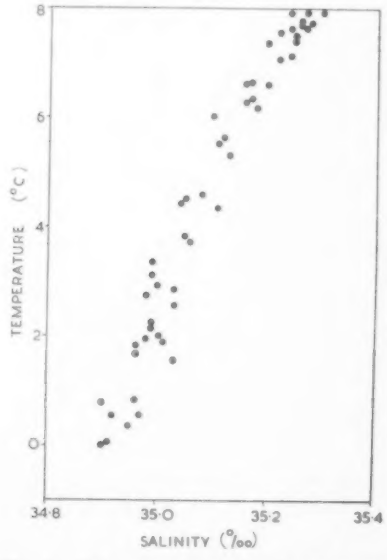

Fig. 3. T-S values in the lower waters on the Ridge.

Vol 6 1959

The main features are shown by the near bottom temperatures in Fig. 1 (depth contours are from DIETRICH, 1956). The deepest reversing bottle on each cast was placed within about 5 m of the bottom by lowering the wire until a slight relaxation in the tension of the 'accumulator spring' on the Kemp davit showed that the lead had touched bottom, and then taking in roughly 3 m of wire. The vertical shape of the overflow on the Atlantic side is given by the profile in Fig. 2 which is taken from the stations joined by the continuous line in Fig. 1. The range of salinities in the lower waters is indicated by the T-S diagram, Fig. 3. Finally the probable influence of this flow in waters to the west is shown by the lower parts of temperature profiles from two stations in deep water, Fig. 4. The lowest 200 m were sampled by reversing bottles spaced at 25 m intervals and show a distinct thermocline in this range. This area was revisited on the 21st and 22nd of September, 1958 but there was no longer any sign of an overflow, Fig. 5. The lowest near-bottom temperature on the Atlantic side was 1·92°C and on the Arctic side, where the temperatures had been 0°C in August, they were above 1°C in September indicating a downward displacement of the colder water.

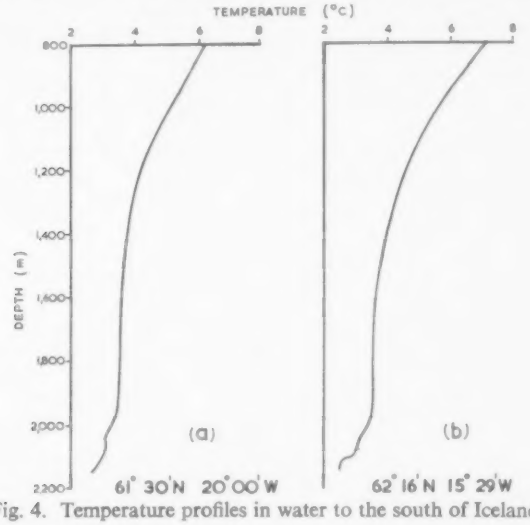

Fig. 4. Temperature profiles in water to the south of Iceland.

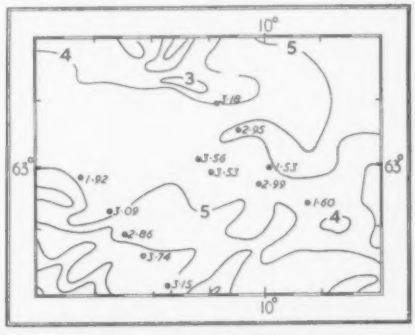

Fig. 5. Bottom temperatures on the Ridge. Sept. 21-22, 1958.

The tentative conclusions from these observations are that the flow, while it occurs, has the form of a continuous stream which spreads laterally but remains in a thin vertical layer on the bottom so that even in the deeper water the flow is not simply geostrophic. These observations also suggest that in future sampling near this area, closely spaced bottles near the bottom would be advisable.

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Absorption curves of acetone extracts of deep water particulate matter

(*Received* 1 *April* 1959)

MEASUREMENTS of light scattering in sea-water have revealed the presence of considerable quantities of particulate matter at depths far below euphotic layers. The size of this particulate matter and its effects on the attenuation of submarine illumination have been extremely controversial subjects in oceanography. Many workers have felt that these particles are of much greater size than wavelengths of light that they scatter and hence they do not selectively scatter any particular visible wavelength. CLARKE and JAMES (1939) observed that the increased attenuation of light in sea-water was caused largely by 'suspendoids' that could be removed by fine filtration. The 'suspendoids,' according to these workers, did not selectively attenuate light. According to Lord Rayleigh if particles are small compared to the wavelength of radiation, scattering will be proportional to $\lambda/4$. This selectivity due to small size causes shorter wavelengths to be more strongly attenuated. UTTERBACK's data (1936) indicates that the selective absorption in sea-water is due, in part, to selective scattering. KALLE (1938) and BURT (1955a, b; 1958) are also of this opinion. KALLE has also shown that sea-water at times contains water soluble pigments which are yellow in colour, and claims that these pigments are representative of some fairly stable metabolic product of phytoplankton.

On a recent summer cruise to the Sargasso Sea (CRAWFORD, No. 17) we were able to collect large quantities of water at depths of 300, 500 and 1000 m. Particulate matter from these depths was immediately concentrated by passing the water through a Millipore-type AA filter. The pore size of this filter is approximately 0·80 μ. After about 10 l. of water had passed through the filter, signs of partial clogging were apparent, and at this point the filter became perceptively coloured. Generally 20 to 25 l. of water from each sample was filtered, after which the filters were removed and stored under refrigeration in the dark. In the laboratory, the filters were extracted in 90 per cent acetone for 24 hours in a darkened refrigerator. Each extract was adjusted to a constant volume of approximately 25 cm^3 and centrifuged in a standard clinical centrifuge for 15 minutes. The absorption spectra of the greenish extracts were measured between 400 and 700 mμ at 5 mμ intervals using a Beckman DU spectrometer with a 5 cm light path.

The absorption spectra of the extracts of deep water particulate matter are characterized by strong absorption between 400 and 500 mμ (Fig. 1). When these curves are compared to absorption curves from extracts of the surface water samples the former differ markedly by either the reduction or the absence of the chlorophyll peak in the region of 665 to 670 mμ. In many respects the shape of the absorption curves of the deep samples suggests the presence of algal carotenoids ; hence one might assume that the deep water extracts are representative of phytoplankton which has spent a considerable period in the darkness. It is well known that light is necessary for the development of chlorophyll *a*, and prolonged periods of darkness generally results in a gradual diminution of this chlorophyll peak.

The ratio of peak absorbencies at 420–430 mμ to 660–670 mμ in natural phytoplankton populations range between 2-3 ; in nitrogen-starved algal cultures this ratio may be as high as four. Absorption ratios of this type for the 0 and 300 m samples are within the natural phytoplankton range; however, the 500 and 1000 m samples have ratios greater than twenty, thus illustrating the extent to which the

greater absorption of blue light occurs in these extracts. If we are to assume that particulate matter at these depths is representative of dead and decomposing phytoplankton it is quite obvious from the absorption curves that the most stable component of the pigment system are carotenoids which absorb most strongly in the blue region.

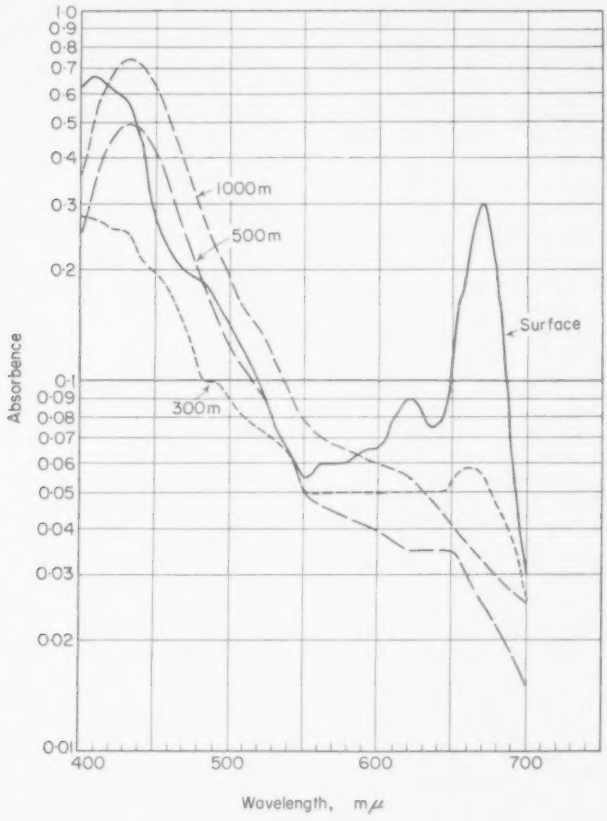

Fig. 1. Absorption curves of acetone extracts of particulate matter from the surface and three depths in the Sargasso Sea. July, 1957.

Regardless of the form of the particulate matter at these depths our observations indicate that the particles are strongly coloured. The actual size of the particles retained by the millipore filter are too large to have any selective influence on the scattering of visible light. However, the colour of these particles and their absorption of the blue light portion of the spectrum exert a selective influence on the attenuation of submarine radiation. This point appears to have been overlooked by many workers. It would appear that many of the observations of the selective attenuation of light by particulate matter may be partly, if not largely, due to absorption by coloured particles rather than being caused by Rayleigh scattering alone.

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Vol
6
1959

BOOK REVIEW

HILARY B. MOORE: **Marine Ecology**. John Wiley, New York, 493 pp.

BECAUSE of its importance to human welfare, research in marine ecology is a rapidly growing science, supported by both civil and military agencies. More and more students are becoming interested in both basic and applied ecology. Unexplored areas in remote parts of the world are being investigated, even in the most inaccessible parts of the Arctic and the deepest trenches of the oceans. Undreamed-of new techniques and methods are developed almost every day permitting new approaches to problems and their solution, which only yesterday seemed insoluble. No wonder that the number of publications in marine ecology has increased to such a degree that today even a specialist will have difficulty in keeping abreast of developments in his own field. It is almost impossible to follow events in related areas, especially as the results are scattered throughout the literature and often difficult to find. The many modern text-books in general ecology devote little space to marine problems. For many years a text-book in marine ecology has therefore been anxiously awaited. Fortunately, in recent years, a series of outstanding books on special topics such as the ecology of the deep-sea, the shore and the open sea have been published, but until recently nobody has had the courage to try to review and co-ordinate into one book all the most important data from the many different branches of marine ecology and related sciences. This is easily understood. To write such a book, which must be more than a mere compilation and especially a book which must cover all aspects of the subject without being too comprehensive and detailed, is an enormous and difficult task. This requires a skilled author with wide interests and experience in different parts of the world and from various fields of this many-faceted science. Moreover the author must not become discouraged because such a book cannot be complete, because many readers will complain that their speciality should have had a more comprehensive treatment, and finally because the book will be in need of revision and additions almost on publication.

This long awaited book on marine ecology has now appeared. The publishers, John Wiley & Sons, could hardly have found a more experienced man than Dr. Hilary B. Moore to write it. As a marine biologist at Millport, Port Erin, Plymouth and Bermuda, and now Assistant Director of the Marine Laboratory of the University of Miami, he has a varied background. The list of references immediately shows that his own interests cover many branches of marine ecology. As might be expected from such an experienced ecologist the result is excellent and the publishers, the author, and not least all marine biologists are to be congratulated on this ideal book. In addition to all its other merits it is also typographically attractive.

Moore's "Marine Ecology" starts with a valuable and interesting chapter on earlier more important books on this and related sciences. Here I miss a reference to the first, German edition of Ekman's "Tiergeographie des Meeres." Although the second, English edition is mentioned, the first edition antedated it by nearly twenty years and exerted a great influence on marine zoogeography. The introduction is followed by a series of chapters on the physical, chemical and biological factors in the environment, which one must consider in solving ecological problems. In the following chapters the chief marine habitats are described and the living conditions in each are surveyed.

The second half of the book is devoted to the organisms inhabiting different habitats, and their reactions to the combined factors of the environment in these habitats. At the end of the book there is a short review, a systematic appendix of species mentioned in the text, an index and no less than 27 pages of references. The list of references is of course far from complete but it is astonishingly up-to-date, and more European books are cited than usual in American text-books.

Though the text comprises only a little more than 400 pages it is remarkably full of information. In spite of this the reader never gets the impression he is reading a dry book of reference. The language is clear and the text should be easily understood even by those who in this book will first become acquainted with marine ecology. It can be read with pleasure from cover to cover by those who are already marine ecologists by profession. The book is profusely illustrated with no less than

243 figures, mostly diagrams, and 72 tables to illustrate the text. Many of the figures chosen are classical. There are not many problems which are not discussed or at least mentioned. If anybody feels that some subject should have been more thoroughly discussed, e.g. the ocean depths and their fauna – and he may be right – the references will give him a hint where to find such information. On the whole, I believe that Dr. Moore has made a wise selection in choosing what should and what should not be considered. His " Marine Ecology " will not only be of interest to the ecologist, but also to the physical or chemical oceanographer, if they want to know how animals and plants react to the factors they are studying. A valuable characteristic of the book is that it not only describes what is known but also clearly points out the problems that are still unsolved. This no doubt will stimulate further research.

It is good to note that Dr. Moore uses the metric system and the Celsius temperature scale, although exceptions occur, where fathoms and Fahrenheit degrees are used. These could readily have been converted to the conventional scales. In some cases, when using tables and illustrations from other sources, a conversion was of course difficult. A table at the end of the book comparing the different units would be of great help. Finally I hope that the next edition will contain fewer misspellings than the first. In fact this contains many, which should have been avoided, starting in the table of contents with " antecology " for 'autecology.' Such errors are, however, obvious. But when authors names, places, genera and book-titles are misspelled this may lead to a repetition of these errors in other publications. In a book such as this, the young student for the first time will encounter the names of many leading biologists. Certainly, then their names ought not to be so maltreated as is the case here with Bruun, Jespersen, Jonsgård, Poulsen and Spärck, only to mention some of the Scandinavians (or even such Americans as Urey, Chace, Irving, etc. Likewise a man should be given his correct initials, e.g. J. L. Fuller, A. J. C. Jensen, F. P. Shepard). Then, too, the well-known " Gullmar " Fjord became ' Gullmer,' " Laccadive " became ' Lacadive,' " Bellinghausen " became ' Bellinghousen,' etc. Hardy's " The Open Sea " was cited as ' The Open Ocean ' ! I hope that this final criticism will not keep anyone from buying and reading this outstanding book. It is really worth reading and useful to all who are interested in the sea and its life. It certainly will take over the role of Murray and Hjort's " The depths of the ocean " for a long time to come.

H. BRATTSTRÖM

Vol 6 1959

Marine geological work of the Soviet Antarctic Expedition, 1955-1957

A. P. LISITZIN and A. V. ZHIVAGO

(*Received* 30 *May*, 1958)

Abstract—In accordance with plans for the International Geophysical Year, the Marine Antarctic Expedition of the U.S.S.R. Academy of Sciences began oceanographic work, especially work in submarine geology, in 1955 in the Antarctic and in the Indian Ocean. Echo soundings in South Polar seas were taken over tens of thousands of miles, particularly in little known areas and new submarine elevations were found in an extensive area of recent volcanic activity west of the submarine mountain range extending between Kerguelen and Gaussberg. For the first time in Antarctic waters, cores up to 16 m long were obtained and samples of bottom sediments were collected on the two voyages at more than 200 stations. A systematic study was made of the distribution and composition of particles suspended in sea-water, as well as the seismo-acoustic work on the thickness of the unconsolidated sediments on the ocean bottom. The Expedition is at present continuing work in the Indian and Pacific Ocean.

Vol.
6
959-60

INTRODUCTION

In accordance with plans for the International Geophysical Year, the Marine Antarctic Expedition of the U.S.S.R. Academy of Sciences was organized in 1955. During the first two years, the Antarctic waters of the Indian Ocean were surveyed and two traverses made of the Indian Ocean (Fig. 1) on the *Ob* (Fig. 2), one from November 30, 1955 to June 6, 1956 and the other from November 7, 1956 to June 21, 1957. A third is in progress. On the first trip the work was largely confined to the southern and central parts of the Indian Ocean and the Tasman Sea. Altogether, 153 stations were made, with geological work at 120 of them. These were taken along the coasts of the eastern Antarctic continent from the Davis Sea to the Balleny Islands, thence over the New Zealand submarine mountain range to the tip of New Zealand, touching *en route* at Macquerie Island. After a short stop at Wellington, a section was made across the Tasman Sea to Sydney and then from Adelaide southward to the Davis Sea, which at the time was mostly covered by ice. The last leg of the journey was a section with 41 complete oceanographic stations from the Antarctic to the Red Sea with a stop at Kerguelen Island.

On the second voyage, several oceanographic sections were made along the stretch from the *Mirny* observatory to Princess Ragnhild Coast, with a few landings on the shores. With an interruption to assist the Japanese Ice-Breaker *Soya*, a section was then run from the Antarctic due north to Capetown, thence to *Mirny* in the general vicinity of the Sub-Antarctic islands. Finally, a section was run along the 97th meridian, stopping only at Cocos Islands, to the mouth of the Ganges River. This completed the oceanographic work. Only soundings and meteorological observations were made on the homeward journey.

Four types of work were undertaken by the submarine geologists on both Antarctic voyages :

1. The topography of the bottom was determined by echo soundings over a

distance of more than 50,000 miles, 20,000 miles being in the little-known regions of the Antarctic.

2. Samples of bottom sediments, many of which were cores 14–16 m long, were collected at 307 stations with various snappers (grabs), core samplers, piston-corers and large-diameter corers.

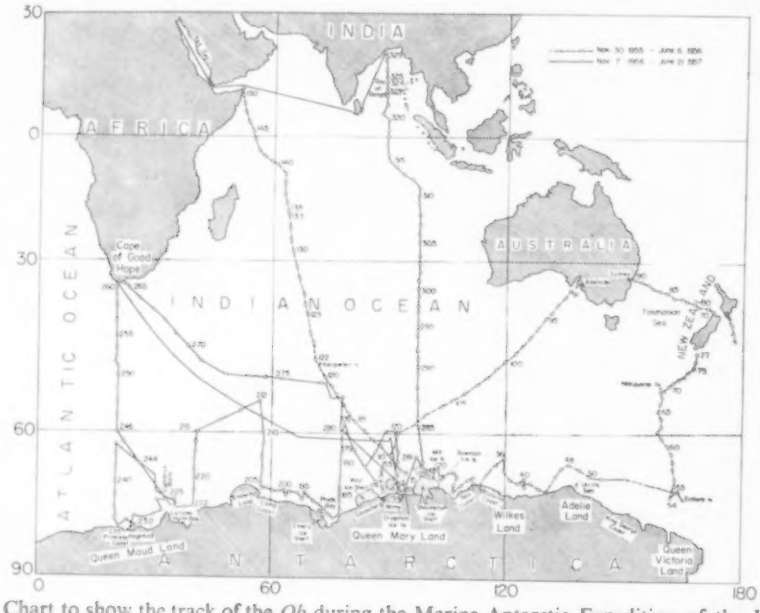

Fig. 1. Chart to show the track of the *Ob* during the Marine Antarctic Expedition of the U.S.S.R. Academy of Sciences, 1955-1957. The tracks between Kaliningrad and the South Polar Observatory 'Mirny' are omitted because no oceanographic work was undertaken on this portion of the voyage, other than soundings on the second trip.

3. Using seismic methods, the reflexions from 250 explosions permitted a determination of the thickness of the unconsolidated sediments.

4. Suspended matter from 6000 samples at the surface and from different depths were collected by membrane filters. In addition, suspended matter has been obtained by pumping 11,000 tons of surface water and 500 tons from deeper water. Large water samplers were also used. Several hundred samples of dust, spores and pollen were collected from the air over the ocean.

LABORATORY FACILITIES AND METHODS

There are four laboratories on the *Ob* for geological work (LISITZIN and ZHIVAGO, 1958c) with a total floor-space of roughly 100 m². These are used as follows : (a) for echo-sounding and seismic work, (b) for bottom sediment work and separation of suspended matter, (c) for filtration and (d) for the microscopic study of sediments and suspended matter. There is also a special drying chamber with two powerful electric driers and a shaking device for the sieve analysis of sands and coarse clastic material, a room for the preparation of thin, transparent sections, as well as storerooms for scientific specimens, scientific equipment, instruments, supplies, etc.

There are four self-recording echo-sounders using dry electrothermic paper. Two use ultrasonic frequencies, two audio-frequencies. The echo-sounders have an automatic timer from a contact chronometer and a device to switch in their receiving and transmitting transducers in parallel, when an

Fig. 2. The diesel-electric ship *Ob* is a modern ship built to withstand the ice, of 12,600 tons displacement, a length of about 130 m, a width of 18 m, and a speed of 15·5 knots. The *Ob* was largely re-equipped for work in the Antarctic. On deck, there are six deep-sea electric 'Ocean' winches with 6-10,000 m of cable, a deep-sea geological winch, with a double drum and 14,000 m of cable and a heavy duty trawling winch with about 11,000 m of cable. Winches were also installed to operate electrical current meters, an electrical thermometer, a bathythermograph, a wave recorder, etc. There are laboratories for meteorology, hydrology, chemistry, biology, marine-geology and geophysics.

Vol
6
1959

especially powerful signal is required. These permit uninterrupted recordings under the most unfavourable conditions, such as during storms or when steaming among ice-floes.

Bottom sediments were obtained with a snapper (' Ocean-50 ') (LISITZIN and UDINTSEV, 1952) sampling an area of 0·25 m^2 and with locking trap doors for operation in rough seas. Hinged doors permit a more rapid lowering of this gear. Samples of sediments 20–25 cm thick have been obtained.

Cores up to 565 cm have been obtained with an ordinary gravity corer of stainless steel. It has a tube of 8 m in length and an inner diameter of 50 mm and weighs 150 to 300 kg. It may be equipped with various cutting tips depending on the type of bottom.

In solid rock, the snapper or ordinary gravity corer does not always work. In such instances, a short gravity corer is used. It has a short tube (about 1 m in length) made of highly resistant steel with pig iron weights up to 300–400 kg. If lowered at maximum speed, it often cuts good cores or brings back stone fragments in the teeth at its tip.

A much simpler corer than Kullenberg's was developed at the Institute of Oceanology. It is made of stainless steel and fits the tube of the gravity corer. This instrument is assembled over the side of the vessel in a vertical position from separate tubes 7-8 m in length using a special Cardan suspension. Separate tubes are connected by bayonet joints.

Some of the geological work requires larger samples than are obtained with the usual corers. Thus, particularly large samples are needed for age determination studies by the ^{14}C and ^{10}Be methods, for determination of the organic content of the sediments, etc. For these purposes, a tube of large diameter was used; this has a more advantageous ratio of cutting surface to inner diameter and a considerably more rigid tube. The percussion tube (' Antarctic ') designed for the expedition had a cross-section of 150 mm and a total length of 13 m. On one occasion a core about 6 m long and weighing 150 kg when wet was obtained in the southern Indian Ocean.

Cores of sands, corals and other compacted sediments were obtained from depths up to 200 m with a vibro-piston corer (KUDINOV, 1957) which is equipped with a high frequency electromechanical vibrator and a piston which allows recovery of undistorted cores. A hydraulic ejector (KUDINOV, 1951) extracts all cores from the tubes. For large-diameter corers, four hydraulic pumps were used simultaneously.

The equipment for a seismic station (' Echo-56 ') includes two pre-amplifiers with piezo-electric hydrophones, three high frequency and three low frequency amplifier channels, filters and recording devices from a loop oscillograph to magnetophones. There are also control and calibrating instruments. Both the refraction and the reflexion methods are possible with this equipment.

Particles of dust from the air were collected by a special installation for membrane ultra-filtration, as well as plates smeared by a sticky substance, which were attached to the ship's bow.

Particles suspended in sea-water which form bottom sediments were studied both from the surface layers and from vertical sections down to a maximal depths. Particles up to 0·7 μ were filtered from 20 to 30 samples at a time on a semi-automatic installation (LISITZIN, 1955 ; 1956).

To obtain larger samples of suspended materials required in the studies of grain size, chemical and mineralogical composition, determination of the organic content and other analyses, a method of separation with powerful disk separators made of stainless steel and other non-corrosive materials treated 150-200 tons of sea-water a day. Water was collected at the surface along the ship's course, but at oceanographic stations the pumps brought up 12 tons of water per hour from depths down to 200 m. The use of four powerful separators in the geological laboratory made it possible to obtain samples of suspended materials from several depths at a station.

On both trips, the separation for the collection of suspended matter from sea-water was effective. Thus, on the second voyage alone, it amounted to 15 kg (dry weight). A study of suspended materials provides reliable data on quantitative distribution in the ocean and on its qualitative composition. For more exact quantitative determinations water samples from the surface were sent through 4-6 membrane ultra-filters and 2–4 separators at the same time.

To obtain 200 l. of water from all depths, a large water sampler was constructed of stainless steel and other non-corrosive materials and was successfully tested at depths of 3000-4000 m. The water from the samplers is transferred through a stainless steel pipe to stainless steel settling tanks of 200 litre capacity, these are required for the ^{14}C method of the absolute age determination of sea-water. Samples of water from various layers can later be driven through smaller separators or suspended matter can be precipitated by coagulation.

In addition, preliminary examination of samples and records is undertaken on shipboard by methods described earlier (LISITZIN and PETELIN, 1956). Determinations of pH and EH of bottom sediments by a lamp potentiometer, microscopic analysis of samples with polarized light and binocular microscopes with artificial light, studies of suspended matter, photography of the wet cores, photomicrography of sediments and separate fractions, sieve analysis, heavy mineral separation in Moshev's ultra-centrifuge, mounting of slides, preparation of thin sections and grinding of rocks and concretions from the sea-bed, study of plant pigments from the sediments and suspended matter by a spectrophoto meter, simple chemical determinations, etc.

In the Antarctic, work is carried out under most unfavourable weather conditions ; in rain or snow, with waves of 10–15 m, in fog, and with frequent ocean swells. With the ship's roll in a high wind, it is difficult to determine when an instrument touches bottom ; water with air bubbles and the vibration of the ship make echo-sounding difficult. Studies are also substantially hampered by ice-floes and icebergs. These may be insurmountable obstacles even for the most powerful icebreakers. Even where a ship may penetrate, it is not always possible to carry on the observational work. Especially difficult are the autumn-winter conditions, when the air temperature drops and severe storms are frequent.

During the second voyage of the *Ob* nearly 100 out of 178 deep-sea stations were made during winds exceeding Force 6 ; under these conditions geological work with heavy gear had to be undertaken with the ship hove-to. In quieter conditions the ship was usually drifting. The geological stations were successfully carried out at the scheduled locations even with a wind up to Force 10–12.

RESULTS

Although the material is still being worked up, it is already possible to draw some conclusions about the geological structure (Figs. 3–7) in this little known part of the Antarctic (ZHIVAGO and LISITZIN, 1957). The sediments and topography of the Davis Sea have now been studied in great detail as the result of numerous crossings of the *Ob* and the *Lena* and from soundings taken by a cutter along ice-free coastal stretches. The Davis Sea has an average depth of about 500 m and with alternating elevations and depressions of 200–300 m. Characteristic is the lateral extension of orographic lines parallel to the continental edge. The littoral is covered by the continental ice out to depths of 80–100 m or more.

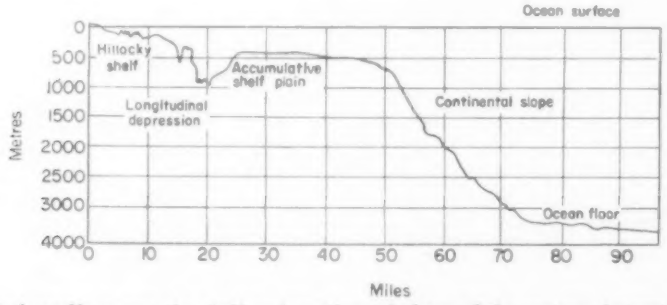

Fig. 3. Typical profile across the shelf and continental slope of the eastern Antarctic continent.

Most characteristic of the bottom topography in the Davis Sea is an elongated depression (Figs. 3–4) on the continental shelf extending eastward under the Shackleton Ice Shelf. In the middle of the Davis Sea this depression is not as deep as to the east. The bottom there does not fall below 790 m with a width between the 700 m isobaths of about 2·5 miles. To the east, depths down to 1400 m and a width of 10–12 miles have been recorded for this depression. Near the Shackleton Ice Shelf, the depression

Vol 6 1959

narrows again to 4–4·5 miles within the 700-m isobaths. On a chart (Fig. 7), it appears as a wide band slightly convex towards the land. At the bottom of the depression, there are long and narrow secondary depressions, which are separated from one another by step-like or level ridges. All are asymmetrical. The sides toward the continent are very steep, while offshore these are gently sloping (Fig. 3). In the western part of the Davis Sea the depression expands and the depths decrease considerably. North of the West Ice Shelf it reaches the depths of the continental slope and cannot be traced farther west.

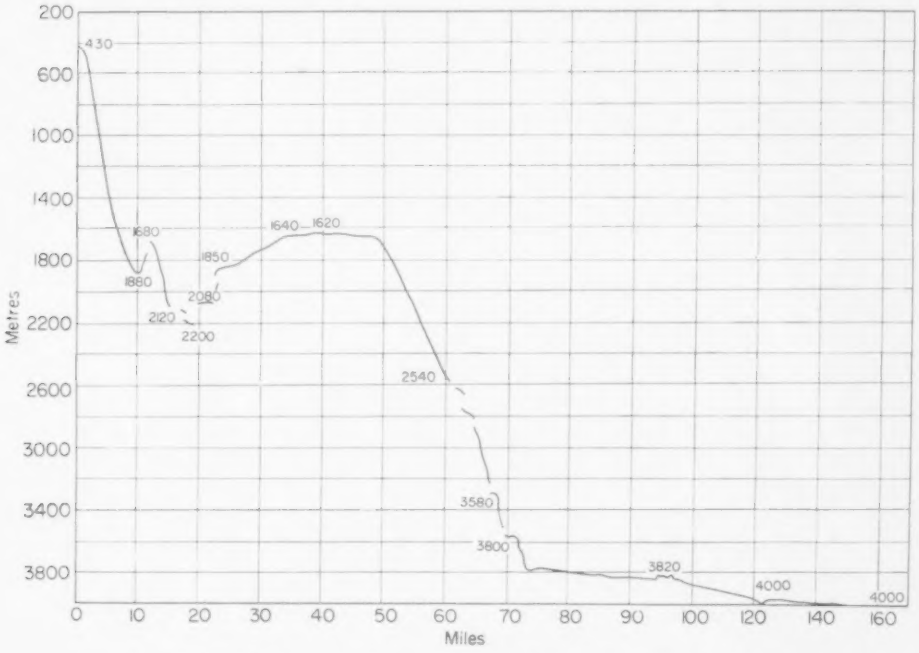

Fig. 4. Section of the continental slope of the Antarctic continent along 97°E.

On both sides of the depression in the Davis Sea are stretches of bottom which are regarded as the Antarctic continental shelf. Average depths are 300-400 m with elevations up to 100-150 m and depressions down to 600 m. The character of the shelf south and north of the depression differs markedly.

South of the depression, up to the coast line, the shelf has numerous elevations and depressions. Within short distances, there are depth differences of 200-300 m. In a number of cases typically glacial forms can be traced on the bottom. Apparently this part of the shelf emerged only comparatively recently from beneath the continental ice and the mountainous country, reworked by glaciers is now partially levelled out by bottom sediments. The topography might be called ' hillocky '.

North of the depression, the shelf is level out to depths of 200-500 m with a flattened or slightly undulating surface. The bottom relief here is much older. Sharp peaks and crests are absent. The level bottom can be correlated with a long and rapid accumulation of terrigenous materials carried out by the ice. The rugged relief of the hillocky shelf is buried under younger sediments.

The upper part of the continental slope of the Davis Sea is a concave scarp descending abruptly down to 2000 m. Farther down, there is a wide slightly tilted step covered by silts and, finally, at the base of the continental slope at depths of from 2400 to 3300 m rounded crests 60-100 m high have been found.

Although the only general features of the topography from the Shackleton Ice Shelf to Balleny Islands have been studied, along one zig-zag line on the first voyage of the *Ob*, the depression on the shelf of the Davis Sea was found apparently extending eastward for more than 2000 miles to d'Urville Sea. The structure of the shelf throughout this expanse is the same as in the Davis Sea (Fig. 3). This regularity in the structure of the shelf could not be accidental and it is, undoubtedly, connected with a comparatively recent regression of the edge of the continental ice; this process is still continuing.

The edge of the shelf east of the Shackleton Ice Shelf, just as in the Davis Sea, lies at a depth of about 400 m, its outer part having a slight tilt towards the ocean, but, the character of the continental shelf east of the Shackleton Ice Shelf changes markedly. Deep crevasses and canyons dissect it into separate blocks, which, on a chart, appear as projections of the main surface of the slope. These projections often have sharp secondary peaks, apparently formed during recent disintegration of rock at the bottom of the slope.

The ocean bed in this part of the Antarctic is characterized by smooth transitions to great depths and a considerable planation. Great depths extend right up to the Balleny Islands with their rocky cliffs rising nearly vertically from the ocean bed. Farther north the base of the islands continues as a submarine ridge which becomes gradually lower and merges with the elevations of the New Zealand submarine mountain range.

As already stated, the Antarctic shelf differs greatly in its morphology and in its structure from that of other continents: its offshore edge sinks to depths of 400-500 m, the bottom at a number of points is extremely irregular with stretches without recent sediments and with a depression on the shelf with depths down to 1000-1400 m extending for a distance of more than 2000 miles.

The width of the shelf varies. Thus, near the Shackleton and West Ice Shelves, it reaches a maximum width of 150 miles, but towards the west and east, it decreases to 30-60 miles. It should be noted that the actual width of the shelf cannot be determined with any accuracy, because the exact position of the coast is hidden beneath the continental ice. For example, the entire territory of Enderby Land could be a vast shelf of ice.

The shelf is usually level at depths of 150-500 m offshore from the longitudinal depression and also west of the Davis Sea – where the depression cannot be traced – on the outer part of the continental shelf. The surface is tilted towards the north, almost entirely smooth and covered by unsorted iceberg sediments. This part of the shelf is apparently older than that adjoining the coast. It has been free from glaciation for a long time and it has long been an area for accumulation of sediments. Small knolls on the shelf are usually due to such an accumulation.

The hillocky portion developed near the coast at depths of 150-200 m can usually be divided into several types determined by : (1) structural features of the bottom, (2) disintegration of rock at the bottom (block shelf), (3) planation by glaciers of bedrock projections and (4) irregularity in the accumulation of iceberg sediments. These

Vol.
6
1959

types are of varying clarity on different parts of the shelf, and preliminary charts of the genetic types have been compiled.

The longitudinal depression on the shelf described for the Davis Sea and the stretch up to Balleny Islands, does not extend beyond the Davis Sea to the West, as shown by soundings during the second voyage. In the vicinity of the West Ice Shelf the depression changes into a wide canyon with a depth down to 700 m and turns north-eastward increasing in depth to that of the upper part of the continental slope. Farther west on the shelf, there are only small separate depressions communicating directly with the great depths of the continental slope. There are, for instance, transverse depressions in the vicinity of Olaf Prydz and Lutzow-Holm bays. It is not known whether these depressions continue under the continental ice sheet. One supposes that they become nearly perpendicular to the shelf and divide it into a number of blocks.

The formation of depressions is apparently associated with vertical movements of the continent during the Quarternary, when large longitudinal and horizontal fractures formed along its periphery. Large fractures are usually accompanied by smaller rifts. In a number of places recent tectonic dislocations and faults have formed blocks on the hillocky shelf. Secondary fractures appear as narrow fissures down to 1000 m which are usually parallel to the coast and with asymmetrical sides not covered by sediments.

At the offshore edge of the shelf, there are wide, sloping mounds, the moraines of ancient glaciers which ended much farther north than those today. These mounds are usually parallel to the coast and the highest ones apparently serve as a base for Bowman, Mill and Drygalski ice islands. The emerged part of these islands is entirely formed of ice. Preliminary geophysical studies confirms the suggestion that the bases formed part of the moraine.

It should be noted that submarine moraines in an area with glacial sediments can be distinguished only by morphological features from sediments deposited by icebergs because the composition of the two types of sediments are very similar.

On the continental slope of the Antarctic continent, there are sloping and undulating step-like plains, as well as stretches divided by large and small longitudinal crests, cross ridges (counter forts) and areas of a block and hillocky dissection.

The sloping plains of the continental slope usually descend from the edge of the shelf and are a direct continuation of the level shelf. In the Shackleton Ice Shelf the sloping plain stretches down to 1500 m. Primary tectonic irregularities are covered by a thick level layer of icebergs sediments.

A younger variety of sloping plain on the shelf is the undulating aggradation plain usually located deeper than the sloping plains. West of Lützow-Holm Bay, for instance, it begins at the edge of the shelf and descend to depths of 1000 m.

The lower part of the Antarctic continental slope in the interval between 2500 and 3700 m is a wide sloping surface divided by short steep scarps. Step-like plains on the continental slope may be traced for great distances, with the exception of an area in the vicinity of 40°E, where a mass of volcanic lava blocks and hummocky surfaces have been found.

In the middle of the continental slope down to depths of 2000 m there often are stretches dissected by large and small longitudinal crests. The height of the large crests reaches 1000 m. Quite often there are 3-4 parallel rows of them. Smaller

secondary longitudinal ridges of 80-100 m occur both on the upper part of the slope and at its base. Often they are correlated with larger crests. Stretches with a perpendicular dissection of the slope to form crossing ridges have been found only in the Kemp coast area. Deeper down the slope are crests which rise up to 1000-1200 m. Apparently they are a continuation of the continental elevation at right angles to the coast. Stretches of block and hillocky dissection of the slope occupy only small areas.

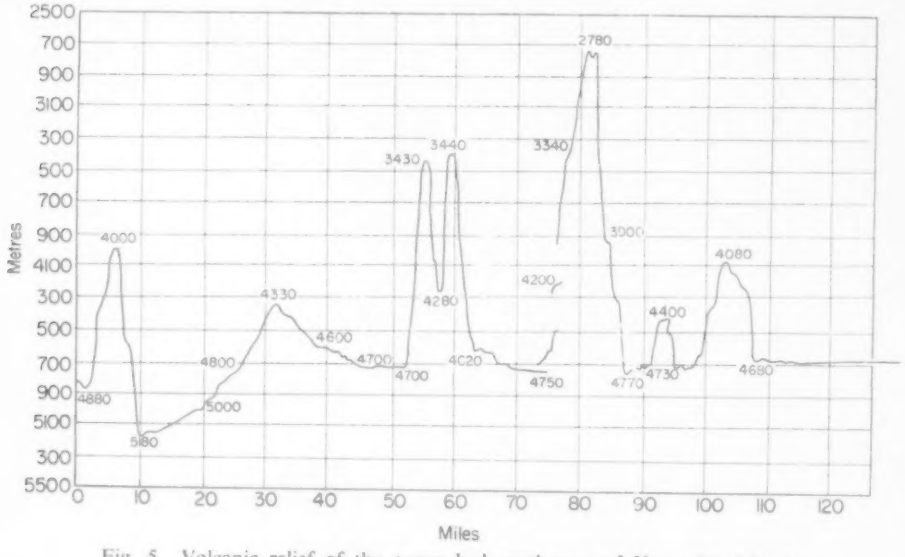

Fig. 5. Volcanic relief of the ocean bed south-west of Kerguelen Island

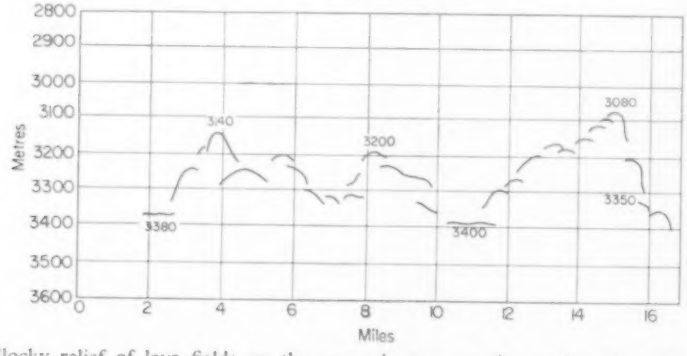

Fig. 6. Hillocky relief of lava fields on the ocean bottom on the section along 97° E, at 45° S.

An extensive area of submarine volcanic activity lies west of the submarine ridge between Kerguelen and Gaussberg. This is apparently a continuation of the volcanic zone found to the eastward in the southern part of the Atlantic. About three-fourths of the bottom of the Indian Ocean between the Cape of Good Hope and the Antarctic west of the Davis Sea is of volcanic origin. Over expanses of the deep-sea bottom are numerous lone volcanoes and clumps of volcanoes (Fig. 5), and vast lava fields, with

Vol 6 1959

step-like block and hummocky surfaces (Fig. 6). Many volcanoes rise more than 3000 m above the bottom. In contrast to the chain of Subantarctic islands, volcanic activity in this area is recent. Over extensive expanses of the sea-bottom there are outcrops of volcanic ashes and lava fields not covered by, or with only a thin layer, of recent sediments on top of them. Pyroclastic material is present even in the surface layers of the sediments. Much volcanic activity took place at the end of the Quaternary period, as proven by numerous bands of ash material in the cores. The section along the 97°E longitude reveals the extent of the volcanic activity in this region. Farther east, there is evidence of recent volcanic activity in the Balleny Islands. In the areas of vulcanism a number of morphological types are delimited. Relatively small areas of the ocean bed are deep-sea plains resulting from the accumulation of sediments, as at the base of the continental slope at depths of 3700-5000 m. This area is covered by iceberg sediments and diatomaceous ooze. This strip tends to parallel the edge of the continent and varies in width from 60 to 150 miles. Only at 90°E does it reach a width of 200 miles.

The eroded undulating relief of the sea-bed is apparently a younger stage in the development of deep-sea plains. A longer accumulation would have completely covered the original tectonic forms. North of Gunnerus Bank there is a limited area of block relief of the volcanic bed with heights of 100-400 m. It is probably associated with fault displacements in this region.

Studies of relief during the two voyages of the *Ob*, in addition to the establishment of genetic and morphological types (Fig. 7) in the Antarctic and southern Indian Ocean (ZHIVAGO and LISITZIN, 1957; LISITZIN and ZHIVAGO, 1958a) afforded an opportunity to trace the coast line in a number of places, to obtain soundings for compiling charts in little known areas, and to locate a number of new submarine elevations.

Soundings in the vicinity of Gribb Bank showed that it is not in the locality printed on the navigational charts. A detailed bathymetric chart in the vicinity of Banzare Bank was also made. New data on bathymetry and sediments were collected from the Kerguelen-Gaussberg and New Zealand ridges.

Two years work on suspended particulate matter in the Antarctic and the Indian Ocean shows that the content of suspended matter in summer in the open sea varies in the surface layer between 0·5 and 0·8 g/m^3, with an increase up to 1·5-1·8 g/m^3 near the coast and near islands. At the edge of the ice along the Antarctic continent, the suspended particulate matter is considerably higher in summer, often about 1-2 g/m^3 and in some places up to 2·8-3·0 g/m^3. This is caused by an abundant growth of diatoms. The convergence zone is precisely marked by an increase of up to 1-2 g/m^3 in the amount of suspended matter in contrast to 0·5-0·8 g/m^3 elsewhere.

The vertical distribution of suspended matter in near-Antarctic regions is complicated. At a number of stations at 25-50 m, concentrations of 6-10 g/m^3 are recorded, but near the continental slope and especially on the upper part, it is 4-5 g/m^3. In the southern and central Indian Ocean at 3000-6000 m, it decreases to 0·1-0·5 g/m^3. However, at certain stations concentrations up to 0·7-1·5 g/m^3 have been recorded. In the Bay of Bengal near the mouth of the Ganges River, the concentration of suspended matter in the subsurface layers is 4-6 g/m^3, but in the surface layers it is considerably lower (0·2-4 g/m^3).

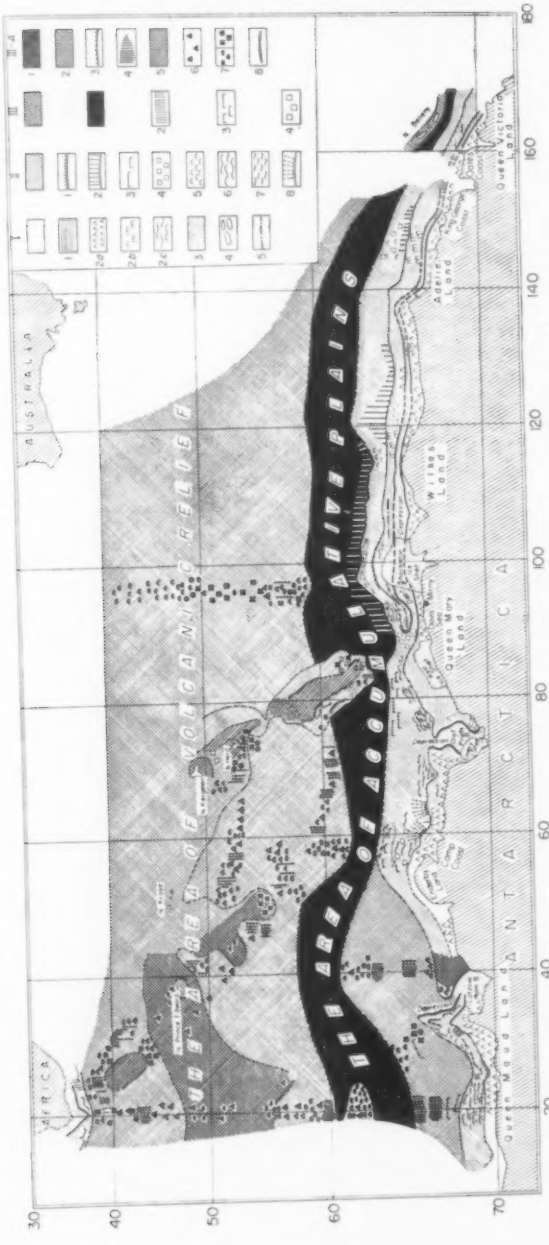

Fig. 7. Bottom relief types of southern Indian Ocean (sketch chart by A. V. ZHIVAGO).

Vol
6
1959

The detailed study of the suspended matter from the membrane filters and separators has not yet been completed, but preliminary examination of the samples with a polarizing microscope showed considerable differences in the composition of the suspended matter from different parts of the ocean and at different depths.

A preliminary study of seismic reflexions reveals several layers of sediments down to depths of 100-1200 m below the surface of the bottom. Data on the thickness and extent of these layers will be reported later. The extent of the layers of volcanic ash in the diatomaceous sediments was traced for great distances with echo-sounders. In a number of cases, 3-4 layers were followed simultaneously for distances of hundreds of miles. Some of these layers were missed by the corers.

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Vol.
6
959-60

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Micro-organisms as indicators of hydrological phenomena in seas and oceans—I

Methods

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Abstract—Current boundaries, the mixing zone of sea and river waters, and the thermocline may be identified by variations in the numbers of the microbial population. In the convergence zone of different water masses, boundary surfaces are formed where organic substances accumulate due to absorption phenomena or for other unknown reasons. As a consequence an increased number of micro-organisms occur there.

Microbiological studies in the Central Arctic showed that even slight variation in the stratification of the water could be detected by determining the abundance of the micro-organisms. The pronounced increase in the microbial population in the boundary layer of waters with only slight hydrographic differences makes it possible to utilize this phenomenon to identify weak, deep ocean currents.

Another method of the utilization of micro-organisms as hydrological indicators has been shown: investigation of areas of microbial forms with peculiar morphological characteristics makes it possible to determine the origin of water masses.

Vol
6
1959

COMPARATIVE microbiological investigations in various parts of the Atlantic Ocean enabled FISCHER (1894) to conclude that the number of bacteria increases considerably at the boundaries of surface currents. He assumed that this phenomenon resulted from the accumulation at the current margins of drifted masses of dead organic material.

BUTKEVICH (1938) also reported an increased number of bacteria in the boundary layer between the warm Atlantic waters and cold waters in the Barents Sea. He believed that such accumulations of bacteria in regions where diverse waters mix results from the death of plankton organisms sensitive to variations in their environmental conditions.

Large populations of micro-organisms are also found in the mixing zone of river water with sea-water often at some distance off the river mouth. Such observations made by BUTKEVICH (1938) and OSNITZKAJA (1954) in the northern part of the Caspian Sea showed a sharp increase in the bacterial concentration of up to 1–2 million per ml. of water, in the region of convergence of the Volga and Caspian waters (Fig. 1). In the Volga effluent and in the superficial layers of the rest of the Caspian Sea, on the other hand, there were only tens or hundreds of thousands of bacteria per ml. BUTKEVICH pointed out that similar phenomena were also observed in the vicinity of the confluence of the Indiga and Tuloma rivers into the Barents Sea and of the Niva and Covda into the White Sea. Thus, not only the high temperature gradient where the warm and cold currents meet but the decrease in salinity, resulting from the mixing of these waters account for the increase of the microbial population in the convergence zone of various water masses. It seems evident that the mixing of waters

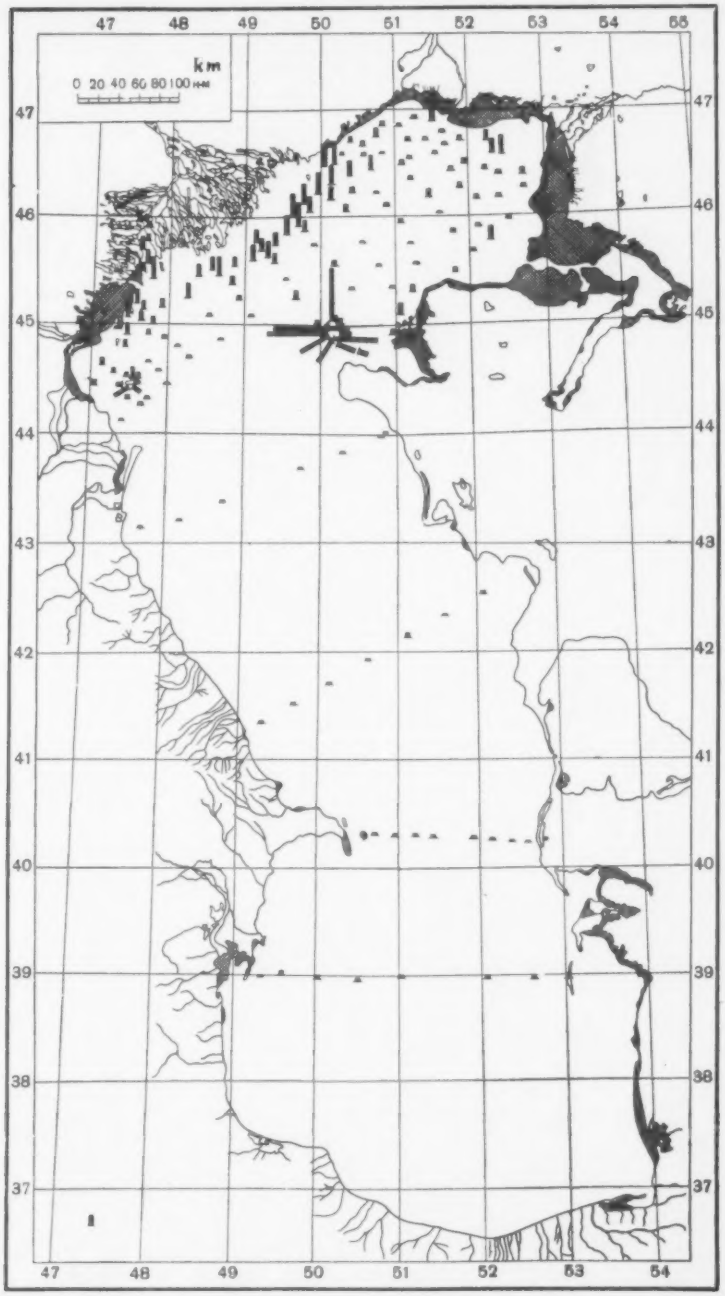

FIG. 1. Number of micro-organisms in the surface layer of the Caspian Sea. The height of the symbol given in the bottom left-hand corner defines the bacterial content. I = One million bacteria per ml. of water.

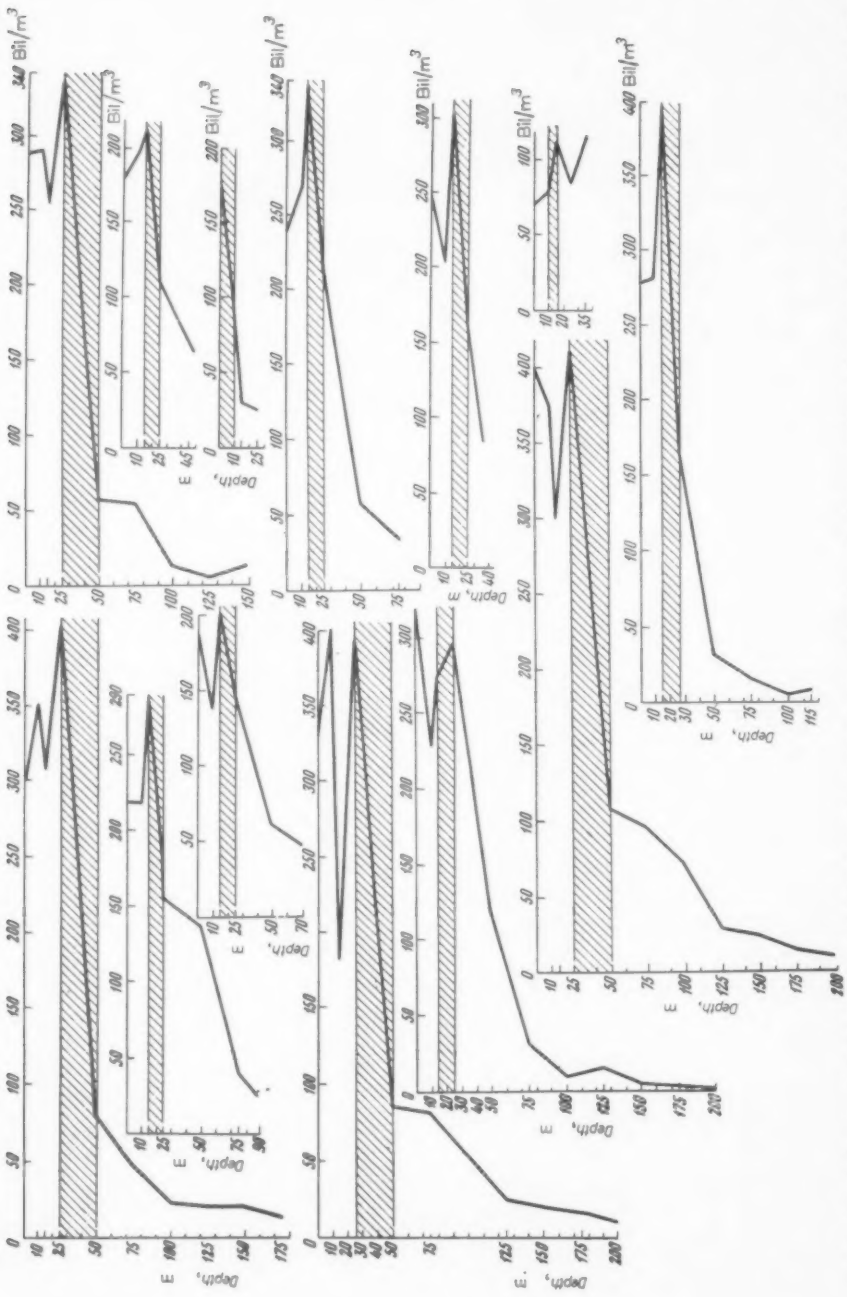

FIG. 2. The influence of the thermocline (shaded) on the quantitative distribution of micro-organisms along vertical sections in the Caspian Sea.

Vol
6
1959

of different densities forms interfaces where organic matter accumulates, and as a consequence, a rapid growth of bacteria takes place there.

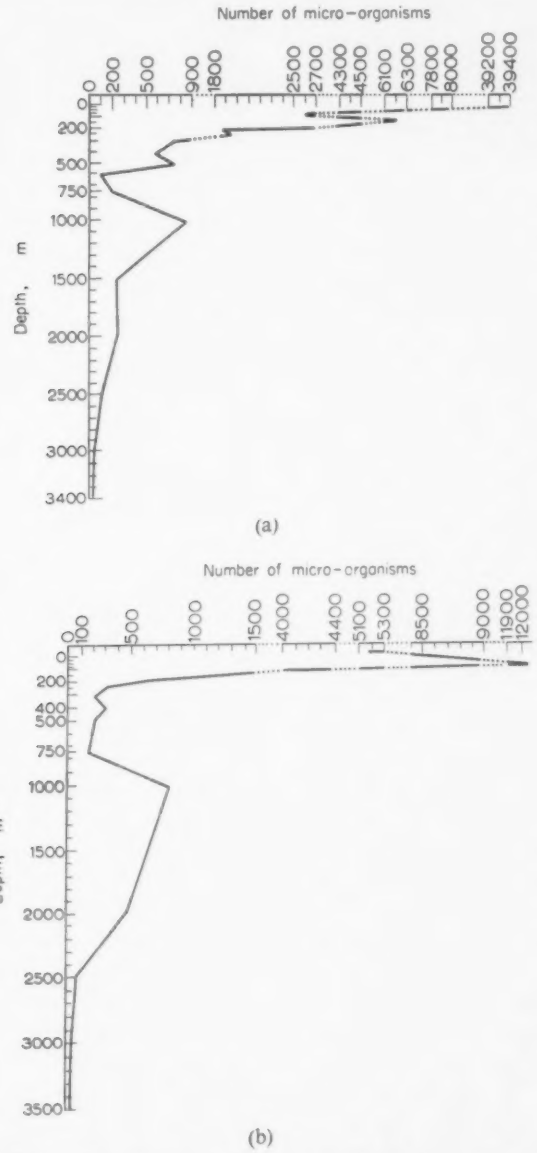

FIG. 3. Vertical distribution of micro-organisms in the region of the North Pole (data obtained by counts on filters) (a) July; (b) September.

We also observed an increase in the microbial biomass in the thermocline (Fig. 2) at all stations in the middle and southern regions of the Caspian Sea where the vertical quantitative distribution of micro-organisms was investigated (KRISS, 1956). This

transition layer between warm surface waters and cold deep waters contributes to an accumulation of the organic substances and, consequently, to a more intensive propagation of microbial cells.

From the above examples, it appears that the variation in the density of the microbial population may provide an indication of certain hydrological phenomena, i.e. in convergence zones of water masses of different origin (boundaries of currents, areas of confluence of the river and sea waters and thermocline).

With conventional methods for counting micro-organisms in sea-water clear-cut differences in the microbial cell concentrations in relation to hydrological factors become obvious. It should, however, be noted that temperature and salinity gradients were great wherever abnormalities in the quantitative distribution of micro-organisms were observed. There were temperatures of several degrees difference between warm and cold currents and in the thermocline; salinity decreases in the ' frontier ' regions between river and sea-waters were no less significant. It was of great interest therefore, to ascertain the degree of sensitivity of the bacteriological method in revealing hydrological inhomogeneities in water masses and the possible application in discovering slight differences in the structure of the sea and ocean waters.

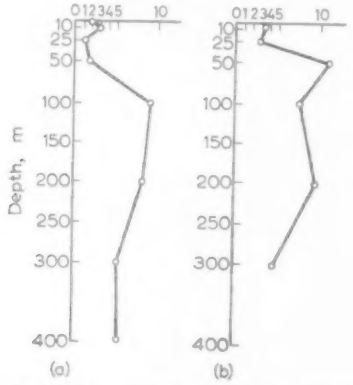

FIG. 4. Vertical distribution of bacteria in the Arctic Ocean (BUTKEVICH, 1938)
(a) *St.* 99 – 82° 42′N and 87° 03′E; total depth – 2365 m.
(b) *St.* 75 – 80° 53′N and 72° 14′E; total depth 550 m.

Investigations of this kind were carried out on an ice-floe research station ' SP-3 ' in the region of the North Pole (KRISS, 1955). As is known, a layer of Atlantic water at depths between 200 m. and 1000 m. occurs in the highest latitudes of the Central Arctic. This intrusion of the Atlantic current into the Polar basin differs from Arctic waters above and below by its warmer temperatures, although these variations were slight, being within a range of 1°C. Thus, the temperature of the Atlantic water was only a few tenths of a degree above zero, whereas that of the Arctic waters was several tenths of a degree below zero. Consequently the density gradient between the Atlantic layer and the adjoining cold waters was insignificant, particularly at the lower boundary.

Microscopic examination of the vertical quantitative distribution of micro-organisms in July, 1954, at the time when the ice-floe station was at 88°N, and again in September when only half a degree from the North Pole, showed a marked increase

of the microbial population in the boundary layer between the upper cold water and the underlying Atlantic water (Fig. 3). A similar increase also occurred at the lower margin of the Atlantic current (1000 m.) and in contrast to layers above and below. It was, however, less dense than those in the upper boundary of this current.

Much farther south in the Arctic Ocean at 81°N and 82°N a similar phenomenon was noted by BUTKEVICH (1938). Thus, direct counts of micro-organisms were higher in the intermediate layer between the Arctic and underlying warm water of the Atlantic current (Fig. 4).

It is noteworthy that slight variations in the density observed in the transition zone of the Atlantic waters and the overlying Arctic waters were also evident from the curve of the vertical quantitative distribution of those bacterial species which represented only a small portion of the microbial population of the water basins and are capable of growth using only easily assimilated organic matter. Such heterotrophic micro-organisms indicated their presence in thousands per litre in water of the transition zone between Arctic and Atlantic water masses, whereas above and below they occurred in hundreds.

The microbiological investigations in the Central Arctic established that even slight differences in stratification are revealed by variations in the quantitative distribution of micro-organisms. In fact, as will be evident from forthcoming papers, the application of this microbiological method increases our knowledge of the hydrological structure of waters in the Indian Ocean and the central part of the Pacific Ocean.

Micro-organisms with morphological peculiarities may also be used as hydrological indicators in establishing the origin of water masses. This was not possible until a technique for direct microscopic examination of the microbial population had been developed, because the morphological structure could hitherto only be ascertained for the comparatively small number of bacterial species which have been grown under laboratory conditions. Due to the scarcity of microbial forms in cultures, chiefly rods and cocci, the morphological characteristics of bacteria could not be relied on as indicators in hydrological studies. An exception is a spore-forming bacteria *Bac. mycoides*; its peculiar colonies easily identify these wide-spread soil bacteria. ISSATCHENKO (1937) found *Bac. mycoides* in parts of the Kara Sea where the Ob-Yenisei waters might be expected. Thus, he considered it possible that the Ob-Yenisei current, extended to 92–93 °E, and then deviated northward, because farther to the east in Vilkitzky Strait no *Bac. mycoides* were found.

Direct microscopical examination makes it possible to observe among the micro-organisms which do not grow on artificial media the forms with unusual morphology (KRISS *et al.* 1958). Some of these micro-organisms are widely distributed in oceans (KRISS and MITZKEVICH, 1957); the habitats of others are comparatively restricted.

Marine microbiology and its relationship with hydrology is still in its early stages, but the discovery of easily distinguished microbial forms associated with definite water masses, suggests the possibility of using such forms as indicators of the origin of water masses and also for the quantitative determination of water exchanges.

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Vol
6
1959

The hadal* community, an introduction

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Abstract—(1) During the last ten years the Danish *Galathea* and the Russian *Vitjaz* Expeditions have made biological investigations in thirteen deep-sea trenches at depths exceeding 6000 m. Due to the complicated configuration of the trenches trawling is often very difficult and can only be undertaken after careful echo sounding.

(2) Environmental features including extent, sediments, pressure, oxygen, salinity, temperature, and food supply were examined.

(3) The quantitative composition of the bottom fauna seems to vary considerably from one trench to another, depending on the abundance of food supply in the photic zone above. Quantitative bottom samples indicate a biomass as high as 10–40 g/m^2 in certain trenches.

(4) In eight of the trenches a total of 250–310 species have been found. The decrease in the number of species with increased depths within these eight trenches is smaller than might be expected (probably at the most 50 per cent). Dominant groups are actinians, polychaetes, isopods, amphipods, gastropods, lamellibranchs, and holothurians. Considering the total number of species in each group, the most important are : Pogonophora, echiuroidean worms, holothurians, and isopods. Decapod crustaceans, brachiopods, and turbellarians have no hadal representatives, and other groups (including fishes and bryozoans) are very insignificant in the hadal zone.

(5) So far 127 species (and subspecies) from depths exceeding 6000 m have been identified. Of these 74 (58 per cent) are endemic to these depths. The upper limit of the hadal zone might better be set at 6800–7000 m resulting in 62 endemic species (43 per cent). As thus defined, fourteen genera and two families (of actinians and pelagic amphipods) seem to be restricted to the hadal zone.

(6) Nine exclusively hadal species are found in more than one trench (two of them in three trenches), but most of these are close to one another geographically.

(7) The communities in the hadal zone are listed in Appendix 3. Several species (especially of holothurians and Pogonophora) must occur in great numbers.

(8) Both the benthic and pelagic hadal animals are without pigmentation and are blind. In probably all crustacean groups living at great depths a remarkable gigantism has been found, probably due to the effect of the hydrostatic pressure on the metabolism.

(9) The hadal fauna is certainly derived from the abyssal zone. Endemic families, genera and perhaps also some species may be considered as relics of a preglacial abyssal and hadal fauna, but the main invasion of eurybathic species came after the onset of the first glaciation.

(10) Very little is so far known of the hadopelagic fauna. The few hauls with closing nets seem, however, to indicate a high percentage of endemism.

INTRODUCTION

THE invention of the echo-sounder some 35 years ago greatly increased our knowledge of the actual depth and the geographical distribution of the maximum deeps of the oceans. The great majority of these deeps occur in the Pacific, are more or less elongated trenches, and located close to small oceanic islands or (more rarely) larger islands or continents (cf. BRUUN, 1957, Fig. 2).

*The following terms, which are essentially synonymous, have been suggested for the zone between about 6000-7000 m and the maximum depths of 11,000 m : *Hadal* and *hadopelagic* (BRUUN, 1956a), '*super-ozeanische Tiefen*' (ZENKEVICH, 1954), and *ultra-abyssal* and *ultra-abyssopelagic* (ZENKEVICH and other Russian authors in later publications).

For a long time nothing was known about the possible existence of the fauna at these maximum depths. Would any organism be able to withstand the great pressures there ? FONTAINE's (1930) experiments on tolerance to pressure made him suggest that the lower limit of life might be found at a pressure of 700-800 atm (i.e. at depths between 7000 and 8000 m).

Only within the last ten years has this question been answered by sampling the fauna from the greatest depths. In 1948 the Swedish Deep-Sea Expedition on the *Albatross* made a haul on the east slope of the Puerto Rico Trench in the Western Atlantic at depths between 7625 and 7900 m (NYBELIN, 1951), 1600-1900 m deeper than that made at 6035 m in 1901 on a cruise of the *Princesse-Alice*. Then, in 1951, the Danish *Galathea* Deep-Sea Expedition first proved that life exists even at the greatest depths with five successful hauls between 9800 and 10,200 m in the Philippine Trench (BRUUN, 1951). Later, four more trenches were investigated (BRUUN, 1953 ; 1956b, WOLFF, 1953). Starting with a single haul in 1949 (at 8100 m in the Kurile-Kamtchatka Trench), the Soviet research vessel *Vitjaz* chiefly between 1953 and 1958 carried out very extensive surveys in several trenches. The results of these are difficult to find, and almost all are published only in the Russian language. In addition to the papers on various animal groups listed below (Appendix 1), there are several very useful reports on trawling in the Kurile-Kamtchatka Trench (ZENKEVICH, *et al.*, 1954, 1955 ; ZENKEVICH, 1954). In more recent results I am only aware of two papers : BIRSTEIN, *et al.* (1958) on trawling in 1957 in the Mariana Trench, and BELJAEV, *et al.* (1958) on trawling in the Tonga Trench, also in 1957. All mention of other work carried out by the *Vitjaz* since 1955, cited in this paper, is derived from two sources : (1) Location, depth, temperature and station numbers for animals from depths exceeding 6000 m, chiefly from the *Trudy Instituta Okeanologii*. (2) Diagrams, illustrations and preserved animals in the Russian deep-sea exhibit during the International Congress of Zoology in London in 1958. I take this opportunity to thank the Russian scientists who arranged this display and allowed me to copy and publish their preliminary details for those interested in deep-sea biology.

These sources of information have made it possible to make a first attempt at a general survey of animal life in the deep-sea trenches and of the hadal community.

TRAWLING TECHNIQUE

In trawling at the greatest depths (KULLENBERG, 1951 ; 1956), it is first necessary to know how much wire to pay out for a given depth at the desired speed of the gear along the bottom to make certain that the trawl just reaches the bottom. The length of wire can be calculated as a function of the wire angle on leaving the ship, the diameter, the resistance of the trawl to the water (depending on the type of gear), and the actual depth. It took 5-6 hrs to pay out the 12 km long steel wire, reaching bottom some 5 km behind the ship. Since most trenches are narrow and uneven in configuration, it is essential to have an echo-sounder recording continuously. Before any trawling can be undertaken there must be days of painstaking sounding to obtain a reliable survey of the bottom topography.

Other difficulties may also be encountered. For instance, during the early trawling of the *Galathea* in the Philippine Trench all factors (wind, current, waves and swell) which served to force the ship off its course came from the east at right angles to the north-south direction of the trench. Since the deepest part of the trench (roughly

Vol 6 1959

10 km) is a narrow plain only 1-3 km wide, it was necessary to head into the current to keep directly over the even bottom, while trawling north or south along the trench. The slopes were so steep that the trawl would be lost if it hit them. Fortunately, the west and east slopes were found to differ so that with the continuously recording echo-sounder, the position of the ship at any given moment could be located – and the course altered accordingly. There are similar problems everywhere in trawling in narrow trenches (i.e. BIRSTEIN, *et al.*, 1958).

THE ENVIRONMENT OF THE DEEP-SEA TRENCHES

The following environmental features seem to be those of greatest importance for life in the trenches.

Extent. Although depths exceeding 6000 m are distributed throughout the oceans, the total area amounts to only 1·2 per cent of the oceans, and is very insignificant compared to areas with depths between 3000 and 6000 m occupying 76 per cent of the oceans and more than half the surface of the globe.

Sediments. So far there has been no thorough geological study of the bottom sediments in the various trenches, but, as in abyssal depths, they are a soft, fine, more or less clayey ooze with oxidized clay and diatom and radiolarian oozes as the main components. Moreover, the presence of large and small jagged stones of littoral origin (BRUUN, 1953, 1956b, p. 181) seems to be a characteristic feature of trenches in areas with frequent earthquakes and eruptions of submarine vulcanoes and are indicative of turbidity currents. These not only supply the deepest water masses with oxygen (and food ?) from above, but also may cause occasional catastrophes to the trench fauna, resulting from mud-slides. Together with stones the great abundance of waterlogged lumps of pumice in southern trenches such as the Kermadec Trench (WOLFF, 1953, Fig. 16) should also be mentioned as a substratum for attached animals such as hydroids, *Stephanoscyphus* and *Scalpellum.*

Pressure. FONTAINE (1930) and several other physiologists have shown that increased pressure has a marked effect on the metabolism and ability to survive in littoral animals which are experimentally submitted to pressures of several hundred atmospheres. These experiments, and especially the demonstration (ZOBELL, 1952 ; ZOBELL and MORITA, 1959) of barophilic bacteria in the deep-sea trenches seem to indicate that organisms from the greatest depths have special physiological adaptations to determine their upper limit of occurrence. Also the lack of certain groups of marine animals (i.e. Crustacea Decapoda) at greater depths is most probably due to the effect of the hydrostatic pressure.

Oxygen. All measurements of the oxygen content in the trenches have shown sufficiently large amounts to support animal life. For instance, the *Galathea* found about 4·5 ml/l in the Kermadec and Tonga Trenches (BRUNN and KIILERICH, 1955). It is not yet known, however, as to whether there is a decrease of oxygen in the water layers immediately above the bottom.

Salinity. The slight differences in salinity so far found in the trenches (34·7‰ ± 0·1‰) is certainly without ecological significance.

Temperature. The bottom temperature of the Aleutian, the Kurile-Kamtchatka, the Riu-Kiu, the Philippine, the Banda, the New Britain, the Kermadec and the Sunda Trenches (Appendix 3) vary between 1·2°C (6200 m in the Kermadec Trench) and 3·6°C (7300 m in the Banda Trench). There is an increase of temperature towards

the bottom, due to adiabatic heating (BRUUN and KIILERICH, 1955, including bibliography).

Food Supply. BRUUN (1956a, 1957) and KRISS (1959) have pointed out that food in the abyss should be considered as any energy stored in organic matter, which can be utilized by heterotrophic bacteria which thus provide an important source of food for many deep-sea animals. On account of the great scarcity of abyssopelagic and hadopelagic life (VINOGRADOV, 1954), the amount of food available in the trenches is probably of the same order of magnitude as in abyssal depths. Both faunas are ultimately dependent on primary production in the photic zone above*. The presence in some trenches of great numbers of Pogonophora whose peculiar mode of extra-oral feeding probably requires a fairly high concentration of food, as well as the extraordinary gigantism of certain hadal animals (see p. 103) suggest that in many trenches the amount of food available may be even greater than at abyssal depths.

THE BOTTOM FAUNA

Quantitative Composition

So far, very few quantitative bottom samplings have been made at depths exceeding 6000 m. In spite of twelve lowerings with a 0·2 m² Petersen grab during the *Galathea* Expedition, only seven were successful, with living animals in five. The deepest sample (the Philippine Trench, 10,120 m) contained only one specimen, a small holothurian, *Myriotrochus bruuni* (weighing 0·1 g). The richest samples were taken in the Banda Trench and contained the following species :

6580 m	7270 m
1 *Ancistrocyllis constricta* (Polych.)	2 *Macrostylis hadalis* (Isop.)
3 *Tharyx multifilis* ,,	1 *Harpinia spärcki* (Amph.)
1 *Travisia profundi* ,,	2 *Cuspidaria* sp. (Lam.)
1 ? *Owenia lobopygidiata* ,,	
1 *Leptanthura hendili* (Tanaid.)	3 species, 5 specimens.
2 *Harpinia spärcki* (Amph.)	Total weight : 2·5 g.
2 *Cuspidaria* sp. (Lam.)	
1 *Ceraplectana trachyderma* (Holoth.)	
8 species, 12 specimens. Total weight : 2·2 g.	

The *Vitjaz* also took bottom samples with a 0·25 m² modified Petersen grab in the hadal zone (Aleutian, Kurile-Kamtchatka and Japan Trenches). A rich sample from the Aleutian Trench at 7246 (7286?) m depth contained a large specimen of the echiurid worm *Vitjazema aleutica*, weighing 10·1 g, and other invetebrates and protozoans weighing 0·1 g. Other bottom samples yielded species of *Macellicephala* and *Macellicephaloides*, *Jakobia birsteini* and two species of *Elpidia*, one of which was *E. glacialis* (ZENKEVICH and FILATOVA, 1958).

During their deep-sea trawling, the Russians attached a special ' trawlograph ' to the sledge trawl (Agassiz trawl) to measure the trawling distance which permits a rough estimate of the quantitative distribution of animals brought up in the trawl.

*Examples of these food sources are : dissolved or colloidal organic matter, the waterlogged plant debris (from huge trunks of trees down to suspended particles), the excrement of animals living between the surface and the bottom, and the enormous number of chitinous integuments shed by many moulting animals, especially crustaceans.

Vol 6 1959

By means of this technique the biomass of both the zooplankton and the benthos of the Kurile-Kamtchatka Trench and the adjacent part of the Pacific has been estimated (ZENKEVICH, *et al.* 1955, ZENKEVICH and BIRSTEIN, 1956). The biomass of the trench between 8330 and 9950 m was found to be 0·3 g/m².

Number of Species

Based on all the literature known to me, on the data of the exhibit in London (mentioned above) and on the *Galathea* collections I have tried to make as comprehensive a list as possible of all animals so far recorded from depths exceeding 6000 m (Appendix 1).

If we exclude all localities with only scattered records of single species (i.e. most Russian records from 1955 onwards), approximately the following numbers of species of the most important major groups were taken from the Kurile-Kamtchatka, Philippine, Banda, New Britain (*Galathea* records only), Kermadec, Sunda, Mariana and Tonga Trenches (in the two latter trenches the deepest trawling only) :

	No. of species		No. of species	
Porifera :		8–10	Pycnogonida :	2
Coelenterata :		23–30	Mollusca :	45–55
Hydrozoa :	4		Gastropoda : 15–20	
Scyphozoa :	1–2		Lamellibranchiata : 25–30	
Anthozoa :	18–23		Bryozoa :	1
Nemertini :		1	Echinoderma :	43–55
Nematodes :		1–3	Asteroidea : 8–10	
Polychaeta :		50–55	Ophiuroidea : 6– 7	
Echiuroidea :		7–9	Holothurioidea : 25–35	
Sipunculoidea :		4–5	Pogonophora :	5
Priapuloidea :		1	Enteropneusta :	1
Crustacea :		58–65	Tunicata :	2
Tanaidacea :	3		Pisces	3–5
Isopoda :	28–30			
Amphipoda :	20–25		Total	250–310

Vertical Distribution within the Hadal Zone

There is less of a decrease in the number of benthic species within the hadal zone (Table 1) than might be expected, but, on the whole, this is very uniform from one group to another. It is only slight for the molluscs and greatest for the echinoderms. The reason for the latter may be the non-hadal status of most crinoids, sea-stars and brittle-stars. A somewhat larger decrease in the number of species seems to occur in the Kurile-Kamtchatka Trench (ZENKEVICH, 1955; ZENKEVICH and BIRSTEIN, 1956) than in the other trenches investigated.

It should be noted that the difficulties in trawling successfully increase with increasing depth and that the risk of animals washing out of the bag or the bucket in the cod end of the trawl on the way up increases with greater depths.

The maximal recorded depths of the various marine groups is given in Appendix 2.

Table 1. *Number of species in each major group between* 6000 *and* 10,700 m, *for intervals of* 1000 m, *based on the same data as in the preceding section*

Depth range (m)	6000–7000	7000–8000	8000–9000	9000–10000	10000–10700
No. of hauls	9	6	6	4	5
	Number of species				
Porifera*	7–8	—	1–2	—	—
Coelenterata	13–14	7	7	4	2
Polychaeta	27	16	17	6	7
'Gephyrea'	6–8	2	4–5	1	1
Crustacea	26	18	12–13	4	6–7
Mollusca	20	13	9–10	9	4
Echinoderma	31	13	10–11	7	5
Pogonophora	—	—	3	3	—
Others	6	5	2	1	1
Total number	136–140	74	65–70	35	26–27
Calculated number, if nine hauls in all	136–140	111	98–105	79	47–49

*BELJAEV *et al.* (1958) record a fragment of skeleton of a hexactennelid sponge from the Tonga Trench (10,687–10,415 m). Since the authors do not, however, include it in the list of the total number of species from this trawling I have also omitted it in Table 1 and in Appendix 2.

Vol 6 1959

Dominant and Insignificant Animal Groups

The dominant groups in the trenches are coelenterates (actinians), polychaetes, crustaceans (isopods and amphipods), molluscs (gastropods and lamellibranchs), and echinoderms (holothurians) (cf. list on p. 99 and Appendix 1).

If the number of species occurring deeper than 6000 m is considered as a function of the total number of species in each major group, the following four groups are the most important : Pogonophora, Echiuroidea, Holothurioidea and Isopoda. Russian scientists have estimated that the following percentages of the total number of species of these same groups are found deeper than 6000 m (Russian exhibit, 1958) : Pogonophora 30 per cent, Echiouroidea 9 per cent, Holothurioidea 1·7 per cent, and Isopoda 1·4 per cent. Deeper than 8000 m the following percentages, respectively, 27, 4, 0·8 and 0·2, are found.

It is also evident (list on p. 99, Appendix 1) that some major groups, which play an important role in littoral, bathyal and even abyssal depths, are totally missing or very insignificant in the hadal zone. Decapod crustaceans, brachipods and turbellarians have never been recorded from depths exceeding 5300 m and fishes, tunicates, cirripeds, bryozoans, and sponges have only a few representatives at depths greater than 6000 m.

Exclusively Hadal Species and Endemism

Although it is premature to draw any definite conclusions, it appears that the true hadal animals (ciphers in bold type in Column 1, Appendix 1 and WOLFF, 1959) only occasionally are found at depths of less than 6800-7000 m. At present, they only total 62 species* (49 per cent) of the 127 identified species (and subspecies)

*When the Russian results are worked out, this number will probably be increased two or three times.

of Metazoa (Appendix 1). If, however, the upper limit of the hadal zone is set at 6000 m (rather than 6800-7000 m), there are 74 such endemic species (58 per cent of the total number of species found deeper than 6000 m).

The number of species which are exclusively hadal (deeper than 6800-7000 m), in the various major groups, already worked up, are as follows (with the percentage of all identified trench species in each group given in brackets) :

Hydrozoa :	2	(67%)	Crinoidea :		
Actiniaria :	5	(84%)	Asteroidea :		
Polychaeta :	8	(36%)	Ophiuroidea :	1	(10%)
			Echinoidea :		
Echiuroidea :	4	(80%)	Holothurioidea :	8	(50%)
Tanaidacea :	2	(67%)	Pogonophora :	3	(50%)
Isopoda :	15	(68%)			
Amphipoda :	9	(56%)			

Earlier it was stressed that isopods (WOLFF, 1956a, p. 150) have a greater number of species restricted to the trenches than other groups of animals represented by quite a number of species in the hadal zone, such as polychaetes and echinoderms, i.e. are more stenobathic*.

One, or probably two, endemic hadal families have been described (Galatheanthemidae Carlgren, 1956 – Actiniaria – 2 species; Vitjazianidae Birstein & Vinogradov, 1955 – Amphipoda – probably exclusively hadopelagic, 1 species†).

Several genera (with the number of species in brackets) are endemic in the hadal zone, namely :

Actiniaria :
Galatheanthemum Carlgren, 1956 [2]
Hadalanthus Carlgren, 1956 [1]

Polychaeta :
Macellicephaloides Ushakov, 1955 [3]

Echiuroidea :
Jakobia Zenkevich, 1958 [1]
Vitjazema Zenkevich, 1958 [2]

Crustacea :
Parascaphocalanus Brodsky, 1955 [1]
Zenkevichiella Brodsky, 1955 [1]
Herpotanais Wolff, 1956 [1]

Bathycallisoma Dahl, 1959 [1]
Bathyschraderia Dahl, 1959 [1]
Princaxelia Dahl, 1959 [1]

Echinoderma :
Vitjazaster Beljaev, 1958 [1]
Hadalothuria Hansen, 1956 [1]

Pisces :
Pseudoliparis Andriashev, 1955 (subgen.) (2)

Two copopod genera *Parascaphocalanus* and *Zenkevitchiella* are probably hadopelagic (see p. 124).

*This general picture has not been altered by later Russian additions to our knowledge of the hadal fauna.

†Taken twice with closing net, once between 6000 and 8500 m and once between 4200 and 7800 m; further information about the hadopelagic fauna is given below (p. 104).

A few more genera are generally hadal, but have a single abyssal or abyssohadal species : *Alomasoma* Zenkevich, 1958, *Hirondellea* Chevreux, 1889, *Bathyspinula* Filatova, 1958 (subgen.), and *Heptabrachia* Ivanov, 1957.

Distribution in the Various Trenches

A certain number of exclusively hadal species occur in more than one trench :

Macellicephala hadalis : Philippine, Banda and Kermadec Tr.
Jakobia birsteini : Kurile-Kamtchatka and Japan Tr.
Vitjazema ultraabyssalis : Kurile-Kamtchatka and Idzu-Bonin Tr.
Storthyngura herculea : Aleutian, Kurile-Kamtchatka and Japan Tr.
Storthyngura vitjazi : Kurile-Kamtchatka and Idzu-Bonin Tr.
Bathycallisoma schellenbergi : Kermadec and ?Puerto Rico Tr.
Hirondellea (Tetronychia) gigas : Kurile-Kamtchatka and Philippine Tr.
Pardaliscoides longicaudatus : Philippine and Kermadec Tr.
Myriotrochus bruuni : Philippine and New Britain Tr.
Careproctus amblystomopsis : Kurile-Kamtchatka and Japan Tr.

Some of these trenches (Aleutian, Kurile-Kamtchatka, Japan and Idzu-Bonin Trenches) are rather close to one another and are connected at depths of about 6000 m (BIRSTEIN, 1958). Others, however, (Philippine and Kermadec Trenches) are widely separated.

It is of utmost interest to ascertain, whether the number of hadal species common to two or more trenches will increase or not when the fauna of the other trenches becomes known.

In considering the distribution of hadal species in the various trenches (WOLFF, 1956a, p. 150) it was noted that isopods and amphipods had a considerably higher number of species in the Kermadec Trench than in all the other trenches investigated by the *Galathea*, whereas the opposite was the case with echinoderms and polychaetes. This dominance of isopods and amphipods in the Kermadec Trench might be explained by the fact that most of the abyssal species from which the hadal species may have originated occur at high latitudes*. Thus, of the trenches considered, the Kermadec Trench is the most easily accessible from the antarctic region. This theory is supported by the recent record of no less than five species of isopods from the deepest part of of the neighbouring Tonga Trench (see Appendix 1).

The Hadal Community

In describing a community it is most important to give a comprehensive list of the species in the particular locality and to ascertain which species act as predators, prey, food competitors, etc. Very little information of this sort is available from the hadal zone. Between 6000 and 11,700 m, only 57 successful† hauls (trawling and bottom sampling) have yet been made, and in the true hadal zone (beyond some 7000 m) the number is reduced to about 38. For completeness I have listed all trawls and bottom samples (Appendix 3) taken at depths greater than 6000 m, which are presently known to me. Data for many of the Russian trawls, etc. have been gathered from several Russian sources. In papers by various authors inconsistencies were found in

*At any rate this appears to be true for the isopods.
†"Successful" here means touching the bottom and bringing up one or more bottom-living animals.

Vol 6 1959

recording the depth and geographical position at the same station or stations. In such cases the most probable or most frequently used figures were entered in Appendix 3.

Reference is made to all animals, so far recorded from each station. In all cases where sufficient data is available, the total number of species and specimens from the station is included. The number of Foraminifera species has been recorded by the Russians in a number of cases. These are subjoined to the number of metazoan species (Appendix 3). The number of specimens of Foraminifera, however, are not included.

The number of metazoan species from each station varies considerably with the depth and more especially from one trench to another. It is largest in the Kurile-Kamtchatka Trench (up to 45) and the Kermadec Trench (up to 31). This almost certainly is due to an especially rich fauna in these two trenches, compared to the others investigated. It should, however, be borne in mind that owing to the bottom configuration trawling is, for instance, much more easily carried through in the Kermadec than in the New Britain Trench.

The number of specimens also varies greatly. In some places the number especially of holothurians must be enormous compared to that of other animals. Thus, in the Sunda Trench at a depth of 7160 m the *Galathea* collected ca. 3000 *Elpidia glacialis sundensis* and at 8200 m in the Kermadec Trench ca. 1800 *E.g.kermadecensis*. Similarly, the *Vitjaz* brought up from 9000 m in the Kurile-Kamtchatka Trench ca. 1000 *Elpidia glacialis* and almost 2000 specimens of the pogonophores *Zenkevitchiana longissima* and *Spirobrachia beklemischevi*.

Two remarkable photographs were taken by H. E. EDGERTON at 7500 m in the Romanche Trench, (EDGERTON, *et al.*, 1957 ; COUSTEAU, 1958). They show – in addition to an easily recognisable brittle-star – bottom animals which probably can be identified as sea-anemones or stalked barnacles as well as amphipods or isopods.

Special Peculiarities of Hadal Animals

Morphologically the hadal species show the same adaptations to life in eternal darkness as many abyssal and even bathyal (and cave-dwelling) species : greyish or whitish colours and in the hadal zone probably always total blindness.

Gigantism in Crustacea was first noted for almost all hadal species of isopods collected by the *Galathea* (WOLFF, 1956a). These isopods are larger than any other representative of the genus to which they belong. This is also true for tanaids (WOLFF, 1956b), the gigantic hadal *Herpotanais kirkegaardi* being the largest tanaid ever recorded. The hadal mysid *Amblyops magna* is the largest species of the genus and a hadal species of *Haplomesus* (isopods) is two-three times longer than the abyssal representatives of this genus (ZENKEVICH and BIRSTEIN, 1956). Large-sized gammarid amphipods were recorded abysso- and hadopelagically (BIRSTEIN and VINOGRADOV, 1958). A correlation was found between the depth and size of the hadal species of the isopod genus *Storthyngura* from the north-west Pacific, but not between their dimensions and temperature (BIRSTEIN, 1957). The great variety in food habits and source of food in groups of crustaceans showing deep water gigantism indicate that also the food factor certainly can be excluded. It is therefore most probable that this characteristic phenomenon is caused by the effect of hydrostatic pressure on the metabolism, as suggested by ZENKEVICH and BIRSTEIN (1956) and BIRSTEIN (1957).

Gigantism has not yet been reported among the hadal representatives of any animal groups other than Crustacea.

Origin of the Hadal Fauna

There is little doubt that hadal animals are derived from the general abyssal fauna which have many features in common.

Since the relatively sudden onset of the Glacial Age must have resulted in a fundamental temperature change in abyssal and hadal depths of from some 10° to below 4°, only the most eurythermic and eurybathic species could survive. The hadal fauna may have two origins : (1) taxons higher than subspecies or perhaps species and of exclusively hadal occurrence are relics of a preglacial abyssal and hadal fauna ; (2) a new invasion of sufficiently eurybathic species has taken place since the beginning of the first glaciation and is probably still going on (BRUUN, 1956a). The hadal representatives of the family Neotanaidae which are probably archaic and which are presently almost entirely abyssal and hadal (WOLFF, 1956b) are examples of the first type. Hadal species of some isopod genera (i.e. *Macrostylis* and *Storthyngura*) whose closest relatives live in cold water at rather high latitudes in both hemispheres, the most probable place of origin for an invasion of the hadal zone (WOLFF, 1956a), are examples of the second type.

Within deep-sea Malacostraca, there is a dominance of ancient groups in the abyssal and hadal regions (BIRSTEIN, 1958). The ancient fauna with seemingly endemic genera and even families, it should be noted, are not in agreement with the generally accepted idea that most deep-sea trenches are recent, geologically speaking (BIRSTEIN, 1958). They can, however, have migrated from the abyssal zone into the trenches after their formation and later have become extinct in the abyssal zone.

THE HADOPELAGIC FAUNA

Thus far, only the *Vitjaz* has operated closing nets below 6000 m. One haul made in 1953 in the Kurile-Kamtchatka Trench between 6000 and 8500 m revealed an astonishingly rich fauna : twenty species of copepods, four species of ostracods and five species of amphipods (Appendix 4). One of the amphipods is moreover the only known representative of a new, probably hadopelagic family, the Vitjazianidae. One or more hauls between 6000 and 8000 m were made in 1957 in the New Britain Trench (VINOGRADOV, 1958) and with species (probably new) of the genera *Parargissa*, *Andaniexis* and *Hyperiopsis*, fairly closely allied to hadopelagic species of the same genera from the Kurile-Kamtchatka Trench (BIRSTEIN, 1958).

Moreover, three abyssohadal hauls with closing nets have been made in the Kurile-Kamtchatka Trench (4190-8050 m, 4200-7800 m, and 4295-6550 m) and the amphipods in them enumerated (BIRSTEIN and VINOGRADOV, 1955).

According to ZENKEVICH and BIRSTEIN (1956) the hadopelagic animals are consistently unpigmented while the abyssal plankton is usually characterized by an intense dark red colour. This seems to indicate vertical migration by the abyssopelagic, but not by hadopelagic animals which are completely isolated from the illuminated water layers above.

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Vol 6 1959

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Vol
6
1959

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Vol.
6
959-60

APPENDIX 1

List of bottom animals from depths exceeding 6000 m, *their number and depth range in the trenches, depth range in other localities and total number of records. Species printed in bold types indicate those known exclusively from depths exceeding* 6000 m (*in a few cases* 5850 m). *The exclusively hadal species and subspecies (from depths exceeding* 6800–7000 m) *also have their cipher (to the extreme left) printed in bold type*.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
1	FORAMINIFERA I (5 spp.)	many	Kermadec	9995–10,002	Exhibit 1958	Vit.	—	—	—
2	II (4 spp.)	< 50	Tonga	10,687-10,415	BELJAEV *et al.*, 1958	,,	—	—	—
3	*Ammobaculites agglutinans* (d'Orbigny)	—	Kur.-Kamt.	6860–7230	STSCHEDRINA, 1958	,,	—	—	—
4	*Ammodiscus* sp.	—	,,	8100	,,	,,	—	—	—
5	*Astroriza* sp.	—	,,	7210–7230	,,	,,	—	—	—
6	*Cyclammina* sp.	—	,,	6860	,,	,,	—	—	—
7	*Glomospira gordialis* (Jones & Parker)	—	,,	8100	,,	,,	—	—	—
8	*Haplophragmoides glomeratus* Brady	—	,,	6860–8100	,,	,,	—	—	—
9	*Haplophragmoides nitidus* (Goës)	—	,,	8100	,,	,,	—	—	—
10	*Haplophragmoides sphaeriloculus* (Cushman)	—	,,	7210–7230	,,	,,	—	—	—
11	*Haplophragmoides* sp.	—	,,	8100	,,	,,	—	—	—
12	*Hormosina globulifera* Brady	—	,,	6860	,,	,,	—	—	—
13	*Hyperammina elongata* Brady	—	,,	7210–7230	,,	,,	—	—	—
14	*Hyperammina* sp.	—	,,	9000–9050	,,	,,	—	—	—
15	*Miliolina* sp.	—	,,	7210–7230	,,	,,	—	—	—
16	*Nodosinum* sp.	—	,,	,,	,,	,,	—	—	—
17	*Recurvoides* sp. I	—	,,	6860–7230	,,	,,	—	—	—
18	*Recurvoides* sp. II	—	,,	9000–9050	,,	,,	—	—	—
19	*Reophax bacillaris* Brady	—	,,	6860–7230	,,	,,	—	—	—
20	*Reophax bilocularis* Flint	—	,,	7210–7230	,,	,,	—	—	—
21	*Reophax nodulosus* Brady	—	,,	6860–9050	,,	,,	—	—	—
22	*Reophax* sp. I	—	,,	7210–7230	,,	,,	—	—	—
23	*Reophax* sp. II	—	,,	8100	,,	,,	—	—	—
24	*Saccammina difflugiformis* Brady	—	,,	,,	,,	,,	—	—	—
25	*Thurammina compressa* Brady	—	,,	9000–9050	,,	,,	—	—	—
26	*Thurammina papillata* Brady	—	,,	,,	,,	,,	—	—	—
27	*Tolypammina* sp.	—	,,	6860–7230	,,	,,	—	—	—
28	*Trochammina* sp.	—	,,	8100	,,	,,	—	—	—
29	*Trochaminella bullata* Höglund	—	,,	,,	,,	,,	—	—	—
30	PORIFERA I (3–4 spp.)	4	Kermadec	6620–7000	BRUNN, 1953	Gal.	—	—	(3)
31	,, II (?)	1	Sunda	6730	,,	,,	—	—	(1)
32	,, III (3–4 spp.)	—	Kur.-Kamt.	6860–8660	ZENK. *et al.*, 1955	Vit.	—	—	(3)
33	,, IV (1 sp.) (?)	—	Tonga	10,687–10,415	BELJAEV *et al.*, 1958	,,	—	—	(1)
34	*Hyalonema* sp.	1	,,	6860	,,	,,	—	—	(1)

*As far as yet unidentified species are concerned, the lowest taxon (genus, family, etc.), which was recorded, has been given in the table. Sometimes only the number of a higher taxon (for instance Porifera) was recorded from a certain locality; in such cases the number of species is given after the name of the group.

Vol. 6 1959

Prof. Dr. Hedgpeth
9
859-60

APPENDIX 1—*contd.*

		No. of specimens in trenches	Trench(es)	Depth (m) (in trench)	Recorded by	Research vessel	In one trench only	Depth range outside trenches	Total no. of localities
	COELENTERATA								
35	HYDROZOA (1 sp.)	—	Tonga	7210–7230	BELJAEV *et al.*, 1958	Vit.	—	—	(1)
36	**Aglaophenia(?) galatheae** Kramp, 1956	—	Sunda	7000–6900	KRAMP, 1956	Gal.	+	—	1
37	*Aglaophenia tenuissima* Bale, 1914	—	Kermadec	6660–6770	,,	,,	—	293–385	3
38	*Cryptolarella* (?) sp.	—	Kur.-Kamt.	6860	ZENK. *et al.*, 1955	Vit.	—	—	(1)
39	**Halisiphonia galatheae** Kramp, 1956	—	Kermadec	8210–8300	KRAMP, 1956	Gal.	+	—	1
40	SCYPHOZOA (1 sp.)	2	Kur.-Kamt.	6860	ZENK. *et al.*, 1955	Vit.	—	—	(1)
41	*Stephanoscyphus simplex* Kirkptr., 1890	c. 160	Banda Kermadec	6490–6650 6180–7000	KRAMP, 1959	Gal.	—	430-5860	38
42	*Stephanoscyphus* sp.	1	,,	9995-10,002	Exhibit., 1958	Vit.	—	—	—
	ANTHOZOA								
43	HEXACORALLIA (1 sp.)	1	Kur.-Kamt.	8610-8660	ZENK. *et al.*, 1955	,,	—	—	(1)
44	ACTINIARIA (2–5 spp.)	c. 10	,,	6860-9050	,,	,,	—	—	(3)
45	,, sp.	13	Kermadec	6660–6770	CARLGREN, 1956	Gal.	—	—	(1)
45	**Bathydactylus kroghi** Carlgren, 1956	12	,,	8210-8300	,,	,,	+	—	1
47	**Daontesia mielchei** Carlgren 1956	1	Banda	7290-7250	,,	,,	+	—	1
48	**Galatheanthemum hadale** Carlgren, 1956	50	Philippine	9820-10,210	,,	,,	+	—	3
49	**Galatheanthemum profundale** Carlgren, 1956	c. 120	Kermadec	5850–8300	,,	,,	+	—	6
50	Galatheanthemidae gen. I (1 sp.)	—	Aleutian	7246	Exhibit., 1958	Vit.	—	—	(1)
51	Galatheanthemidae gen. II (1 sp.)	2	Kur.-Kamt.	7210-7230	ZENK. *et al.*, 1955	,,	—	—	(1)
52	Galatheanthemidae gen. III (1 sp.)	—	Japan	6156–6570	Exhibit. 1958	,,	—	—	(2)
53	Galatheanthemidae gen. IV (1 sp.)	—	Mariana	10,630-10,710	,,	,,	—	—	(1)
54	**Hadalanthus knudseni** Carlgren, 1956	3	Kermadec	6660-6770	CARLGREN, 1956	Gal.	+	—	1
55	**Paredwardsia lemchei** Carlgren, 1956	40	Sunda	7160	,,	,,	+	—	1
56	CERIANTHARIA (2–3 spp.)	3	Kur.-Kamt.	6860-7230	ZENK. *et al.*, 1955	Vit.	—	—	(2)
57	ANTIPATHARIA (1 sp.)	2	Sunda	7000-6900	BRUUN, 1953	Gal.	—	—	1
58	*Bathypathes patula* Brook, 1889	2	Kur.-Kamt.	8632-8779	PASTERNAK, 1958	Vit.	—	1331–5303	9
59	OCTOCORALLIA (1 sp.)	1	,,	8610–8660	ZENK. *et al.*, 1955	,,	—	—	(1)
60	*Umbellula* sp. (1–2 spp.)	2	Kermadec	6180–6730	BRUUN, 1953	Gal.	—	—	(2)
61	NEMERTINI (1 sp.)	1	Kur.-Kamt.	7210-7230	ZENK. *et al.*, 1955	Vit.	—	—	(1)
62	NEMATODES I (1 sp.)	1	Tonga	10,687–10,415	BELJAEV *et al.*, 1958	,,	—	—	(1)
63	,, II	12	Kermadec	9995–10,002	Exhibit., 1958	,,	—	—	(1)
		7	,,	6620-6730	BRUUN, 1953	Gal.	—	—	(1)
64	POLYCHAETA I (9–12 spp.)	—	Kur.-Kamt.	6810-9050	ZENK. *et al.*, 1955	Vit.	—	—	(3)
65	,, II (1 sp.)	1	Tonga	10,687-10,415	BELJAEV *et al.*, 1958	,,	—	—	(1)
66	,, III (2 spp.)	many	New Britain	8980-9043	Exhibit. 1958	,,	—	—	(1)
		1	,,	8940	—	Gal.	—	—	(1)
67	**Ammotrypane galatheae** Kirkegaard, 1956	6	Banda	7290–7250	KIRKEGAARD, 1956	,,	+	—	(1)
68	*Ammotrypane* sp.	—	New Britain	8980–9043	Exhibit., 1958	Vit.	—	—	(1)
69	Ampharetidae gen. I (2–3 spp.)	c. 60	Kur.-Kamt.	7210-8430	ZENK. *et al.*, 1955	,,	—	—	(2)
70	Ampharetidae gen. II (1 sp.)	1	Sunda	7000–6900	KIRKEGAARD, 1956	Gal.	—	—	(1)
71	Ampharetidae gen. III (1 sp.)	1	Kermadec	6660–6770	,,	,,	—	—	(1)
72	*Amphicteis mederi* Annenkova, 1929	c. 20	Kur.-Kamt.	8100	USHAKOV, 1950	Vit.	—	88–1366	2

APPENDIX 1—*contd.*

		No. of specimens in trenches	Trench(es)	Depth (m) (in trench)	Recorded by	Research vessel	In one trench only	Depth range outside trenches	Total no. of localites
73	*Ancistrocyllis constricta* Southern, 1921	1	Banda	6580	KIRKEGAARD, 1956	Gal.	—	c. 5	2
74	Cirratulidae (*Cossura* ?)	2	Kermadec	9995–10,002	Exhibit. 1958	Vit.	—	—	(1)
75	Cirratulidae gen. (1 sp.)	1	Tonga	10,687–10,415	,,	,,	—	—	(1)
76	Hesionidae (1 sp.)	4	New Britain	8980–9043	,,	,,	—	—	(1)
77	*Ilyphagus bythincola* Chamberlin, 1919	1	Sunda	6730	KIRKEGAARD, 1956	Gal.	—	3436	2
78	*Jasmineira* sp. I	2	Banda	7240–7290	,,	,,	—	—	(1)
79	*Jasmineira* sp. II	1	Kermadec	6620–8300	,,	,,	—	—	(2)
80	*Kesun abyssorum* Monro, 1930	1	Sunda	6730	,,	,,	—	300–1470	6
		11	Kermadec	6960–8300					
81	*Lumbriconereis* sp.	4	Kur.-Kamt.	8100	USHAKOV, 1952	Vit.	—	—	(1)
82	*Macellicephala abyssicola* Fauvel, 1913	—	Aleutian	7246	Exhibit., 1958	,,	—	4255–4872	9
		1	Banda	7290–7250	KIRKEGAARD, 1956	Gal.			
		31	Kermadec	6620–7000	,,	,,			
		2	Sunda	7160	,,				
83	**Macellicephala hadalis** Kirkegaard, 1956	2	Banda	7290–7250	,,	,,	+	—	5
		36	Kermadec	6660–8300					
		2	Philippine	10,150–10,210					
84	*Macellicephala mirabilis* McIntosh, 1885	—	Kur.-Kamt.	7210–7230	ZENK. *et al.*, 1955	Vit.	—	13–1366	12
85	**Macellicephala zenkevichi** Ushakov, 1955	1	,,	8100	USHAKOV, 1955	,,	+	—	1
86	*Macellicephala* sp. I	—	Mariana	10,630–10,710	Exhibit., 1958	,,	—	—	(1)
87	*Macellicephala* sp. II	1	Pto. Rico	7625–7900	ELIASON, 1951	Alb.	—	—	(1)
88	**Macellicephaloides grandicirra** Ushakov, 1955	c. 90	Kur.-Kamt.	8100–9950	USHAKOV, 1955	Vit.	+	—	3
89	**Macellicephaloides verrucosa** Ushakov, 1055	3	,,	7210–7230	,,	,,	+	—	1
90	**Macellicephaloides vitiazi** Ushakov, 1955	2	,,	7210–8430	,,	,,	+	—	2
91	*Maldanella harai* (Izuka, 1902)	49	Kermadec	6620–6770	KIRKEGAARD, 1956	Gal.	—	146–4700	13
92	Maldanidae gen. (1 sp.)	—	Kur.-Kamt.	7210–7230	ZENK. *et al.*, 1956	Vit.	—	—	(1)
93	*Myriochele* sp.	37	Kermadec	6180–8300	KIRKEGAARD, 1956	Gal.	—	—	(3)
94	*Nephthys elamellata* Eliason, 1951	10	,,	6180–7000	,,	,,	—	4255–4600	6
95	Nephthyidae gen. (1 sp.)	4	New Britain	8980–9043	Exhibit. 1958	Vit.	—	—	(1)
96	**Nereis profundi** Kirkegaard, 1956	9	Banda	7290–7250	KIRKEGAARD, 1956	Gal.	+	—	1
97	*Ninoë fusca* Moore, 1911	18	Kermadec	6620–7000	,,	,,	—	4060	4
98	*Notomastus* sp.	5	,,	8210–8300	,,	,,	—	—	(1)
99	*Owenia lobopygidiata* (?) Ushakov, 1950	many	Banda	6490–6650	,,	,,	—	110–1366	3
100	*Petaloproctus cirratus* (?) Monro, 1937	4	,,	7240–7290	,,	,,	—	36–1046	5
101	Polynoidae gen. (1 sp.)	1	Tonga	10,687–10,415	Exhibit., 1958	Vit.	—	—	(1)
102	*Potamilla* sp.	3	Kur.-Kamt.	8100	USHAKOV, 1952	,,	—	—	(1)
103	*Poecilochaetus* sp.	1	Tonga	10,687–10,415	Exhibit., 1958	,,	—	—	(1)
104	Sabellidae gen. (1 sp.)	—	Kur.-Kamt.	7210–7230	ZENK. *et al.*, 1955	,,	—	—	(1)
105	Serpulidae gen. (1 sp.)	1	Kermadec	6620–6730	KIRKEGAARD, 1956	Gal.	—	—	(1)
106	*Terebellides eurystethus* Chamberlin, 1919	1	,,	6660–6770	,,	,,	—	1410–4750	3
107	Terebellidae gen. (1 sp.)	2	Pto. Rico	7625–7900	ELIASON, 1951	Alb.	—	—	(1)
108	*Tharyx multifilis* Moore 1909	3	Banda	6580	KIRKEGAARD, 1956	Gal.	—	1–590	7
109	*Travisia profundi* Chamberlin, 1919	5	,,	6490–7290	,,	,,	—	975–4064	6

VOL 6 1959

APPENDIX 1—*contd.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
110	*Travisia* sp.	65	Kur.-Kamt.	6860–8430	Zenk. *et al.*, 1955	Vit.	—	—	(3)
111	**Vitiazia dogieli** Ushakov, 1953	1	,,	8100	Ushakov, 1953	,,	+	—	1
112	ECHIUROIDEA I (1 sp.)	—	Philippine	10,150–10,210	Bruun, 1953	Gal.	—	—	(1)
113	,, II (1–3 spp.)	—	Kermadec	6660–8300	—	,,	—	—	(3)
114	**Alomasoma chaetifera** Zenkevich, 1958	2	Aleutian	7246	Zenkevich, 1958a	Vit.	+	—	1
115	*Alomasoma nordpacifica* Zenkevich, 1958	2	Kur.-Kamt.	6860–8430	,,	,,	—	520–4560	9
		1	Aleutian	7246					
		1	Japan	7587					
116	**Jakobia birsteini** Zenkevich, 1958	1	Kur.-Kamt.	8100	,,	,,	+	—	3
		24	Japan	6156–6570					
117	**Vitjazema aleutica** Zenkevich, 1958	2	Aleutian	7246	,,	,,	+	—	(1)
118	**Vitjazema ultraabyssalis** Zenkevich, 1958	50	Kur.-Kamt.	7210–9950	,,	,,	—	5560	7
		5	Idzu–Bonin	9715–9735					
119	SIPUNCULOIDEA (2–3 spp.)	c. 170	Kur.–Kamt.	6860	Zenk. *et al.*, 1955	,,	—	—	(1)
120	*Phascolion lutense* Selenka, 1885	150	Japan	6156–6207	Murina, 1957	,,	—	3568–5830	13
		160	Kur.-Kamt.	6860					
		600	NW Pac.	6272–6282					
121	*Phascolion pacificium* Murina 1957	2	Kur.-Kamt.	6860	Murina, 1957	,,	—	5345	3
		19	Japan	6156–6207					
122	PRIAPULOIDEA (1 sp.)	—	,,	7587	Exhibit., 1958	,,	—	—	(1)
123	*Priapulus* sp.	1	Kur.-Kamt.	7210–7230	Zenk. *et al.*, 1955	,,	—	—	(1)
	ARTHROPODA								
	CRUSTACEA								
124	COPEPODA HARPACTICOIDEA	—	Kermadec	9995–10,002	Exhibit., 1958	Vit.	—	—	(1)
125	OSTRACODA	—	New Britain	7657 (?)	,,	,,	—	—	(1)
	CIRRIPEDIA								
126	*Scalpellum* (*Arcoscalpellum*) *japonicum* Hoek, 1883	1	Kur.-Kamt.	6860	Tarasov and Zevina, 1957	,,	—	508–5500	10
		2	Japan	6156–6207					
		1	NW Pac.	6096					
127	*Scalpellum* sp.	10	Kermadec	6620–7000	Bruun, 1953	Gal.	—	—	(3)
128	CUMACEA	—	New Britain	8006	Exhibit., 1958	Vit.	—	—	(1)
129	*Diastylis* sp.	4	Sunda	7160	Bruun, 1953	Gal.	—	—	(1)
130	TANAIDACEA	—	Kermadec	8928–9174	Exhibit., 1958	Vit.	—	—	(1)
131	**Apseudes galatheae** Wolff, 1956	9	,,	6660–6770	Wolff, 1956	Gal.	+	—	1
132	**Herpotanais kirkegaardi** Wolff, 1956	39	,,	6960–7000	,,	,,	+	—	1
133	**Neotanais serratispinosus hadalis** Wolff 1956	5	,,	6960–8300	,,	,,	+	—	2
134	ISOPODA I (5 spp.)	c. 10	Tonga	10,687–10,415	Beljaev *et al.*, 1958	Vit.	—	—	(1)
135	,, II	3	Kermadec	9995–10,002	Exhibit., 1958	,,	—	—	(1)
136	,, III	8	New Britain	8980–9043	,,	,,	—	—	(1)
137	**Antarcturus ultraabyssalis** Birstein, n. sp.	—	Kur.-Kamt.	7210–7230	,,	,,	+	—	1
138	Bathyopsurus nybelini Nordenstam, 1955	1	Pto. Rico	7625–7900	Nordenstam, 1955	Alb.	+	5500–5987	3

APPENDIX 1—*contd.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
139	**Eurycope galatheae** Wolff, 1956	2	Kermadec	6960–7000	WOLFF, 1956	Gal.	+	—	1
140	**Eurycope madseni** Wolff, 1956	1	,,	,,	,,	,,	+	—	1
141	**Eurycope menziesi** Wolff, n. sp.	1	,,	,,	—	,,	+	—	1
142	**Haplomesus giganteus** Birstein, n. sp.	—	Kur.-Kamt.	8330–8430	Exhibit., 1958	Vit.	+	—	1
143	*Hydroniscus* n. sp.	—	Japan	6475–6570	,,	,,	+	—	(1)
144	**Ilyarachna kermadecensis** Wolff, n. sp.	7	Kermadec	6660–7000	—	Gal.	+	—	2
145	*Ilyarachna* sp.	—	Kur.-Kamt.	8330–8430	Exhibit., 1958	Vit.	—	—	(1)
146	**Ischnomesus bruuni** Wolff 1956	1	Kermadec	6960–7000	WOLFF, 1956	Gal.	+	—	1
147	**Ischnomesus spärcki** Wolff, 1956	2	,,	6660–7000	,,	,,	+	—	2
148	**Janirella macrura** Birstein, n. sp.	—	Kur.-Kamt.	7210–8430	Exhibit., 1958	Vit.	+	—	2
149	**Leptanthura hendili** Wolff, 1956	1	Banda	6580	WOLFF, 1956	Gal.	+	—	1
150	**Macrostylis galatheae** Wolff, 1956	3	Philippine	9820–10,000	,,	,,	+	—	1
151	**Macrostylis hadalis** Wolff, 1956	2	Banda	7270	,,	,,	+	—	1
152	*Storthyngura benti* Wolff, 1956	28	Kermadec	5230–7000	,,	,,	+	—	5
153	**Storthyngura bicornis** Birstein, 1957	150	Japan	6156–6207	BIRSTEIN, 1957	Vit.	+	—	1
154	*Storthyngura chelata* Birstein, 1957	3	Kur.-Kamt.	6860	,,	,,	—	5345	3
		13	Japan	6475–6570					
155	*Storthyngura furcata* Wolff, 1956	7	Kermadec	5850–6770	WOLFF, 1956	Gal.	+	—	3
156	**Storthyngura herculea** Birstein, 1957	5	Aleutian	7246	BIRSTEIN, 1957	Vit.	+	—	5
		11	Kur.-Kamt.	7210–8100					
		13	Japan	6475–7280					
157	**Storthyngura pulchra kermadecensis** Wolff, n. subsp.	7	Kermadec	6620–6730	—	Gal.	+	—	1
158	**Storthyngura tenuispinis kurilica** Birstein, 1957	1	Kur.-Kamt.	7210–7230	BIRSTEIN, 1957	Vit.	+	—	1
159	**Storthyngura ten. tenuispinis** Birstein, 1957	2	Aleutian	7246	,,	,,	+	—	1
160	**Storthyngura vitjazi** Birstein, 1957	15	Kur.-Kamt.	8330–8430	,,	,,	+	—	2
		11	Idzu-Bonin	7305–7315					
161	AMPHIPODA I (5–7 spp.)	c. 20	Kur.-Kamt.	7210–8660	ZENK. *et al.*, 1955	,,	—	—	(4)
162	,, II (1 sp.)	1–2	Tonga	10,687–10,415	Exhibit., 1958	,,	—	—	(1)
163	**Bathycallisoma schellenbergi** Dahl, 1959	1	Kermadec	6960–7000	DAHL, 1959	Gal.	+	—	2
		5–10	Pto. Rico (?)	7625–7900	SCHELLENBERG, 1955	Alb.			
167	*Bathyceradocus stephenseni* Pirlot, 1934	1	Banda	7290–7250	DAHL, 1959	Gal.	—	1165–1264	2
165	**Bathyschraderia magnifica** Dahl, 1959	3	Kermadec	6960–7000	,,	,,	+	—	1
166	**Eusirus bathybius** Schellenberg, 1955	2	Pto. Rico	7625–7900	SCHELLENBERG, 1955	Alb.	+	—	1
167	**Harpinia spärcfli** Dahl, 1959	3	Banda	6580–7270	DAHL, 1959	Gal.	+	—	2
168	**Hirondellea dubia** Dahl, 1959	1	Kermadec	7640	,,	,,	+	—	1
169	**Hirondellea (Tetronychia) gigas**, Bst. & Vinogr., 1955	4	Philippine	10,020–10,190	,,	,,	+	Also hadopelagic	6
170	**Lepechinella wolffi** Dahl, 1959	1	Kermadec	6660–6770	,,	,,	+	—	1
171	*Metandania islandica* Stephensen, 1925	1	Pto. Rico	7625–7900	SCHELLENBERG, 1955	Alb.	—	1318	2
172	*Onesimoides cavimanus* Pirlot, 1933	1	Banda	6490–6650	DAHL, 1959	Gal.	—	1158	2
173	*Orchomenella abyssorum* (Stebbing, 1888)	1	Kermadec	8210–8300	,,	,,	—	1414–4330	9
174	**Pardaliscoides longicaudatus** Dahl, 1959	1	Philippine	9820–10,000	,,	,,	+	—	2
		1	Kermadec	6180					
175	Phoxocephalidae (1 sp.)	1	Tonga	10,687–10415	BELJAEV *et al.*, 1958	Vit.	—	—	(1)

Vol. 6, 1959

Vol. 6
1959–60

APPENDIX 1—*contd.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
176	Pontogeneidae (?) (1 sp.)	5	Tonga	10,687–10,415	Beljaev *et al.*, 1958	Vit.	—	—	(1)
177	**Princaxelia abyssalis** Dahl, 1959	6	Kermadec	6620–8300	Dahl, 1959	Gal.	+	—	3
178	**Rachotropis flemmingi** Dahl, 1959	5	Sunda	7160	,,	,,	+	—	1
179	*Rachotropis* (?) sp.	1	Kermadec	6960–7000	,,	,,	—	—	(1)
180	**Schisturella galatheae** Dahl, 1959	1	,,	,,	,,	,,	+	—	1
181	**Tryphosa bruuni** Dahl ,1959	2	,,	6660–6770	,,	,,	+	—	1
	MYSIDACEA								
182	**Amplyops magna** Birstein and Tchindonova, 1958	1	Kur.–Kamt.	7210–7230	Birstein and Tchindonova, 1958	Vit.	+	—	1
	PYCNOGONIDA								
183	*Nymphon femorale* Fage, 1956	1	Banda	6490–6650	Fage, 1956	Gal.	—	4040–4330	3
184	*Nymphon profundum* Hilton, 1942	1	Kur.–Kamt.	6860	Zenk. *et al.*, 1955	Vit.	—	3231	2
	MOLLUSCA								
185	LORICATA	—	New Britain	7657	Exhibit., 1958	,,	—	—	(1)
	SOLENOGASTRES								
186	*Neomenia* (?) sp.	1	Kermadec	6660–6770	—	Gal.	—	—	(1)
	GASTROPODA								
187	PROSOBRANCHIA I (6–10 spp.)	152	Kur.–Kamt.	7210–9050	Exhibit., 1958	Vit.	—	—	(3)
188	,, II (1 sp.)	several	New Britain	8980–9043	,,	,,	—	—	(1)
189	,, III (1–2 spp.)	7*	Tonga	10,687–10,415	Beljaev *et al.*, 1958	,,	—	—	(1)
190	,, IV (1 sp.)	3	Kermadec	9995–10,002	Exhibit., 1958	,,	—	—	(1)
191	*Admete* sp.	1	,,	6660–6770	,,	Gal.	—	—	(1)
192	Buccinidae gen. (1–3 spp.)	10	Kur.–Kamt.	6860	Zenk *et al.*, 1955	Vit.	—	—	(1)
193	Natacidae gen. (1–3 spp.)		,,	,,	,,	,,	—	—	(1)
194	*Odostomia* sp. I	1	Kermadec	6660–6770	—	Gal.	—	—	(1)
195	*Odostomia* (?) sp. II	1	,,	8210–8300	Bruun, 1953	,,	—	—	(1)
196	Trochiidae gen. I (1 sp.)	1	,,	6620–6730	—	,,	—	—	(1)
197	Trochiidae gen. II (1 sp.)	9	,,	5850–6770	—	,,	—	—	(2)
198	LAMELLIBRANCHIATA (3–5 spp.)	c. 110	Kur.–Kamt.	8330–9050	Zenk. *et al.*, 1955	Vit.	—	—	(3)
199	*Axinulus* sp. I	21	Tonga	10,687–10,415	Exhibit., 1958	,,	—	—	(1)
200	*Axinulus* sp. II	25	Kermadec	9995–10,002	,,	,,	—	—	(1)
201	*Bathynella fiora* (Dall, 1916)	—	Japan	6156–6207	,,	,,	—	2870	2
202	*Cuspidaria* sp.	4	Banda	6580–7290	Bruun, 1957	Gal.	—	—	(2)
203	*Glomus* sp.	6	Philippine	9820–10,190	Bruun, 1953	,,	—	—	(2)
204	*Hyalopecten* sp.	1	Kur.–Kamt.	8100	Zenk. *et al.*, 1955	Vit.	—	—	(1)
205	*Kelliella* sp. I (cf. *pacifica* E. A. Smith, 1885)	13	Kermadec	6620–8300	Bruun, 1957	Gal.	—	—	(2)
206	*Kelliella* sp. II	11	Sunda	7000–6900	—	,,	—	—	(1)
207	**Ledella ultraabyssalis** Filatova, n. sp.	many	Kur.–Kamt.	9000–9050	Exhibit., 1958	Vit.	+	—	(1)

*One egg capsule included.

APPENDIX 1—*contd.*

		No. of specimens in trenches	Trench(es)	Depth (m) (in trench)	Recorded by	Research vessel	In one tr. only	Depth range outside trenches	Total no. of localities
208	Ledidae gen. (1 sp.)	—	,,	6860–8430	ZENK. *et al.*, 1955	Vit.	—	—	(2)
209	Lucinidae gen. (1 sp.)	2	,,	7210–7230	,,	,,	—	—	(1)
210	*Malletia* sp.	2	Kermadec	6960–8300	—	Gal.	—	—	(6)
211	*Neilonella subovata* Verrill & Bush, 1897	—	Aleutian	7246	Exhibit., 1958	Vit.	—	3830–4540	(1)
212	*Neilonella* sp. I	390	Kur.–Kamt.	7210–7230	ZENK. *et al.*, 1955	,,	—	—	(1)
213	*Neilonella* sp. II		,,	,,	,,	,,	—	—	(1)
214	*Nucula* sp.	4	Kermadec	6660–6770	—	Gal.	—	—	(1)
215	Nuculidae gen. (1 sp.)	4	,,	6960–7000	—	,,	—	—	(1)
216	Pectinidae gen. (1 sp.)	3	,,	6620–7000	BRUUN, 1957	,,	—	—	(2)
217	*Propeamussium* sp.	—	Kur.–Kamt.	6860–7230	ZENK. *et al.*, 1955	Vit.	—	—	(2)
218	**Spinula (Bathyspinula) bogorovi** Filatova, 1958	3	Riu–Kiu	6810	FILATOVA, 1958	,,	—	—	(1)
219	*Spinula* (*Bathysp.*) *oceanica* Filatova, 1958	56	Japan	6156–6207	,,	,,	—	4640–5876	16
		1	NW Pac.	6272–6282					
220	**Spinula (Bathysp.) vitiazi** Filatova, n. sp.	—	Japan	7587	Exhibit., 1958	,,	—	—	—
221	*Spinula* (*Spinula*) *calcar* (Dall, 1908)	4	NW Pac.	6096	FILATOVA, 1958	,,	—	4160–5790	6
222	*Teredo* sp.	c. 50	Banda	7290–7250	BRUUN, 1957	Gal.	—	—	(1)
223	*Thyasira* sp.	125	Kermadec	9995–10,002	Exhibit., 1958	Vit.	—	—	(1)
224	*Xylohaga* sp. I	1	,,	6660–6770	—	Gal.	—	—	(1)
225	*Xylophaga* sp. II	1	Banda	7290–7250	—	,,	—	—	(1)
226	*Yoldia* sp.	1	Kermadec	8210–8300	—	,,	—	—	(1)
	SCAPHOPODA								
227	*Siphonodentalium* sp.	1	Sunda	7000–6900	BRUUN, 1953	,,	—	—	(1)
	CEPHALOPODA								
228	Octopodidae gen. (1 sp.)	1	Kur.–Kamt.	8100	ZENK. *et al.*, 1955	Vit.	—	—	(1)
	BRYOZOA								
229	*Bugula* sp.	—	Kermadec	8210–8300	BRUUN, 1953	Gal.	—	—	(1)
	ECHINODERMA								
230	CRINOIDEA I (1 sp.)	—	Idzu–Bonin	9715–9735	Exhibit., 1958	Vit.	—	—	(1)
231	,, II (1 sp.)	—	Kur.–Kamt.	6860	ZENK. *et al.*, 1955	,,	—	—	(1)
232	*Bathycrinus australis* (A. H. Clark, 1907)	4	Kermadec	8210–8300	GISLÉN, 1956	Gal.	—	2510–4640	5
233	*Bathycrinus* sp.	—	Kur.–Kamt.	9000–9050	ZENK. *et al.*, 1955	Vit.	—	—	(1)
234	ASTEROIDEA (1 sp.)	—	Mariana	7585–7615	Exhibit., 1958	,,	—	—	(1)
235	Brisingidae gen. (2–3 spp.)	—	Kur.–Kamt.	6860	ZENK. *et al.*, 1955	,,	—	—	(1)
236	*Eremicaster vicinus* (Ludwig, 1907)	5	Sunda	7000–6900	MADSEN *in litt.*	Gal.	—	5200–5800	6
		3	Kermadec	6620–6730	,,	,,			
		54	Kur.–Kamt.	6860	ZENK. *et al.*, 1955	Vit.			
237	*Freyella mortenseni* Madsen, 1956	3	Kermadec	5850–6180	MADSEN, 1956	Gal.	—	—	2
238	**Hymenaster blegvadi** Madsen, 1956	38	,,	6660–6770	,,	,,	—	—	1
239	***Hyphalaster*** sp.	—	NW Pac.	6272–6282	Exhibit., 1958	Vit.	—	—	(1)

VOL. 6 1959-

APPENDIX 1—*contd.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
240	*Porcellanaster caeruleus* Wyv. Thomson 1877	1	E. Atlantic	6035	Koehler, 1909	Pr.–A.	—	1435–5600	23
241	Pterastidae gen. (2–3 spp.)	—	Kur.–Kamt.	6860–7230	Zenk. *et al.*, 1955	Vit.	—	—	(2)
242	*Styracaster* n. sp.	—	NW Pac.	6272–6282	Exhibit., 1958	,,	—	—	(1)
243	**Vitjazaster djakonovi** Beljaev, n. sp.	—	Kur.–Kamt.	6860	,,	,,	+	—	(1)
244	OPHIUROIDEA I (2–3 spp.)	38	,,	6860–7230	Zenk. *et al.*, 1955	,,	—	—	(2)
245	,, II (1 sp.)	—	New Britain	8006 (?)	Exhibit., 1958	,,	—	—	(1)
246	,, III (1 sp.)	1	Kermadec	6620–6730	Madsen, 1956	Gal.	—	—	(1)
247	*Amphiophiura bullata* (Wyv. Thomson, 1877)	3	E. Atlantic	6035	Koehler, 1909	Pr.–A.	—	2270–5610	(3)
248	**Ophiacantha opercularis** (Koehler, 1909)	1	,,	,,	,,	,,	+	—	1
249	*Ophiura loveni* (Lyman, 1878)	177	Kermadec	6660–6770	Madsen, 1956	Gal.	—	2515–4755	4
	ECHINOIDEA								
250	*Portalesia* sp. (?) *ourorae* Koehler, 1926	—	Banda	7290–7250	,,	,,	—	440–1690	4
251	HOLOTHURIOIDEA I (5–10 spp.)	—	Kur.–Kamt.	6860–9950	Zenk. *et al.*, 1955	Vit.	—	—	(4)
252	II (1 sp.)	1	Kermadec	9995–10,002	Exhibit., 1958	,,	—	—	(1)
253	Apoda (1 sp.)	2	Mariana	10,630–10,710	,,	,,	—	—	(1)
254	*Benthodytes sanguinolenta* Théel, 1882	8	Banda	6490–7290	Hansen, 1956	Gal.	—	768–5206	29
255	*Ceraplectana trachyderma* H. L. Clark, 1908	6	,,	6490–6650	,,	,,	—	3188–5173	4
256	**Elpidia glacialis kermadecensis** Hansen, 1956	c. 1800	Kermadec	6620–8300	,,	,,	+	—	2
257	**Elpidia glacialis solomonensis** Hansen, 1956	77	New Britain	8780–8940	,,	,,	+	—	2
258	**Elpidia glacialis sundensis** Hansen, 1956	c. 3000	Sunda	6900–7160	,,	,,	+	—	2
259	*Elpidia glacialis* Théel, 1876	—	Aleutian	6410	Exhibit., 1958	Vit.	—	70–4750	c. 20
		c. 1550	Kur.–Kamt.	8100–9950	Zenk. *et al.*, 1955	,,	—		
		—	Japan	6156–7587	Exhibit., 1958	,,	—		
		—	Idzu–Bonin	9715–9735	,,	,,	—		
		28	New Britain	8980–9043	,,	,,	—		
260	*Elpidia* sp.	1	Kur.–Kamt.	8610–8660	Zenk. *et al.*, 1955	,,	—	—	(1)
261	*Euphronides verrucosa* Ludwig, 1894	4	Banda	6490–7290	Hansen, 1956	Gal.	—	2404–4478	7
262	Gephyrothuriidae gen. I (1 sp.)	14	New Britain	8980–9043	Exhibit., 1958	Vit.	—	—	(1)
263	Gephyrothuriidae gen. II (1 sp.)	1	Mariana	10,630–10,710	,,	,,	—	—	(1)
264	**Hadalothuria wolffi** Hansen, 1856	14	New Britain	8780–8940	Hansen, 1956	Gal.	+	—	2
265	*Mesothuria murrayi* (Théel, 1886)	1	Banda	6490–6650	,,	,,	—	254–2516	5
266	**Myriotrochus bruuni** Hansen, 1956	101	Philippine	10,120–10,210	,,	,,	+	—	4
		3	New Britain	8940					
267	*Myriotrochus* sp.	5	Tonga	10,637–10,415	Exhibit., 1958	Vit.	—	—	(1)
268	**Paroriza grevei** Hansen, 1956	18	Banda	6490–7290	Hansen, 1956	Gal.	+	—	3
269	**Peniagone vedeli** Hansen, 1956	c. 1030	Kermadec	6180–8300	,,	,,	+	—	5
270	*Peniagone* (? *Scotoanassa*) sp.	20	Pto. Rico	7625–7900	Madsen, 1953	Alb.	—	—	(1)
271	*Periamma naresi* (Théel, 1882)	114	Sunda	7160	Hansen, 1956	Gal.	—	3292	2
272	*Pseudostichopus villosus* Théel, 1886	35	Kermadec	6660–7000	,,	,,	—	896–5307	9
273	*Pseudostichopus* sp.	3	Kur.–Kamt.	8100	Zenk. *et al.*, 1955	Vit.	—	—	(1)
274	**Scotoplanes galatheae** Hansen, 1956	1	Philippine	9820–10,000	Hansen, 1956	Gal.	+	—	1
275	*Scotoplanes globosa* Théel, 1879	49	Kermadec	6180–6770	,,	,,	—	545–4809	11

APPENDIX 1—*contd.*

		No. of specimens in trenches	*Trench(es)*	*Depth* (m) *(in trench)*	*Recorded by*	*Research vessel*	*In one trench only*	*Depth range outside trenches*	*Total no. of localities*
276	*Scotoplanes* sp.	—	Kur.–Kamt.	6860	ZENK. *et al.*, 1955	Vit.	—	—	(1)
	POGONOPHORA								
277	**Heptabrachia abyssicola** Ivanov, 1952	2	,,	8100	IVANOV, 1952	,,	+	—	1
278	**Heptabrachia subtilis** Ivanov, 1957	2	Idzu–Bonin	9600 (?)	IVANOV, 1957	,,	+	—	1
279	**Lamellisabella Johanssoni** Ivanov, 1957	9	Japan	6156–6207	,,	,,	+	—	1
280	*Siboglinum caulleryi* Ivanov, 1957	—	Kur.–Kamt.	8100	,,	,,	—	23–3400	c. 20
281	**Spirobrachia beklemischevi** Ivanov, 1957	many	,,	9000–9050	,,	,,	—	—	1
282	*Zenkevitchiana longissima* Ivanov, 1957	,,	,,	8330–9950	,,	,,	—	4850– ?	c. 5
283	ENTEROPNEUSTA (1 sp.)	12	,,	8100	ZENK. *et al.*, 1955	,,	—	—	(1)
	TUNICATA								
284	*Cnemidocarpa bythia* (Herdman, 1882)	7	Kermadec	6180–7000	MILLAR, 1959	Gal.	—	—	
285	*Culeolus* sp.	1	Kur.–Kamt.	7210–7230	ZENK. *et al.*, 1955	Vit.	—	—	(4)
		—	Japan	6150–6207	Exhibit., 1958				
		—	NW Pac.	6096–6282	,,				
	PISCES								
286	*Bassogigas profundissimus* (Roule, 1913)	1	E. Atlantic	6035	ROULE, 1913	Pr.–A.	—	5610–5860	3
287	*Bassogigas* sp.	1	Sunda	7160	BRUUN, 1953	Gal.	—	—	(1)
288	**Careproctus (Pseudoliparis) amblystomopsis** Andr., 1955	1	Kur.–Kamt.	7210–7230	ANDRIASHEV, 1955	Vit.	—	—	3
		—	Japan	6156–7587	RASS, 1958a, 1958b				
289	*Careproctus* n. sp.	—	,,	7587	RASS, 1958b	,,	—	—	(1)
290	Liparidae gen. (1 sp.)	5	Kermadec	6660–6770	BRUUN, 1953	Gal.	—	—	(1)

VOL. 6 1959

APPENDIX 2

Maximal recorded depth of major groups of marine animals and of deepest identified species (beyond 6000 m) in each group

Group	*Deepest identified species*	*Depth* (m)	*Locality*	*Research vessel*	*Year*	*Recorded by*
FORAMINIFERA	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
	Rheophax nodulosus Brady	9000–9050	Kur.–Kamt.	Vit.	1953	STSCHEDRINA, 1958
	Thurammina compressa Brady	9000–9050	Kur.–Kamt.	Vit.	1953	STSCHEDRINA, 1958
PORIFERA	—	8610–8660	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955
HYDROZOA	—	8210–8300	Kermadec	Gal.	1952	KRAMP, 1956
	Halisiphonia galatheae Kramp	8210–8300	Kermadec	Gal.	1952	KRAMP, 1956
SCYPHOZOA	—	9995–10,002	Kermadec	Vit.	1958	Exhibit., 1958
	Stephanoscyphus simplex Kirkpatrick	6960–7000	Kermadec	Gal.	1952	KRAMP, 1958
HEXACORALLIA	—	10,630–10,710	Mariana	Vit.	1958	Exhibit., 1958
	Galatheanthemum hadale Carlgren	10,150–10,210	Philippine	Gal.	1951	CARLGREN, 1956
OCTOCORALLIA	—	8610–8660	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955
NEMERTINI	—	7210–7230	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955
NEMATODES	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
POLYCHAETA	—	10,630–10,710	Mariana	Vit.	1958	Exhibit., 1958
	Macellicephala hadalis Kirkegaard	10,150–10,210	Philippine	Gal.	1951	KIRKEGAARD, 1956
ECHIUROIDEA	—	10,150–10,210	Philippine	Gal.	1951	BRUUN, 1953
SIPUNCULOIDEA	—	6860	Kur.–Kamt.	Vit.	1953	MURINA, 1957
	Phascolion pacificum Murina	6860	Kur.–Kamt.	Vit.	1953	MURINA, 1957
PRIAPULOIDEA	—	7587	Japan	Vit.	1957	Exhibit., 1958
BRYOZOA	—	5850	Kermadec	Gal.	1952	—
COPEPODA (HARPACT.)	—	9995–10,002	Kermadec	Vit.	1958	Exhibit., 1958
OSTRACODA	—	7657 (?)	New Britain	Vit.	1957	Exhibit., 1958
CIRRIPEDIA	—	6960–7000	Kermadec	Gal.	1952	BRUUN, 1953
	Scalpellum japonicum Hoek	6860	Kur.–Kamt.	Vit.	1953	TAR. and ZEV., 1957
CUMACEA	—	8006	New Britain	Vit.	1957	Exhibit., 1958
TANAIDACEA	—	8928–9174	Kermadec	Vit.	1958	Exhibit., 1958
	Neotanias serratispinosus hadalis Wolff	8210–8300	Kermadec	Gal.	1952	WOLFF, 1956
ISOPODA	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
	Macrostylis galatheae Wolff	9820–10,000	Philippine	Gal.	1951	WOLFF, 1956
AMPHIPODA	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
	Hirondellea gigas Birstein & Vinogradova	10,190	Philippine	Gal.	1951	DAHL, 1959
MYSIDACEA	—	7210–7230	Kur.–Kamt.	Vit.	1953	BIRST. and TCHIND., 1958
	Amplyops magna Birstein & Tchindonova	7210–7230	Kur.–Kamt.	Vit.	1953	BIRST. and TCHIND., 1958
DECAPODA (CARIDEA)	—	5240–5160	Celebes Sea	Gal.	1951	—
(GALATHEIDEA)	—	5240–5160	Celebes Sea	Gal.	1951	—
PYCNOGONIDA	—	6860	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955
	Nymphon profundum Hilton	6860	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955

APPENDIX 2—*contd.*

Group	*Deepest identified species*	*Depth* (m)	*Locality*	*Research vessel*	*Year*	*Recorded by*
LORICATA	—	7657	New Britain	Vit.	1957	Exhibit., 1958
SOLENOGASTRES	*Neomenia* sp.	6660–6770	Kermadec	Gal.	1952	—
GASTROPODA	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
LAMELLIBRANCHIATA	—	10,687–10,415	Tonga	Vit.	1957	Exhibit., 1958
	Spinula (*Bathysp.*) *bogorovi* Filatova	6810	Riu-Kiu	Vit.	1955	FILATOVA, 1958
SCAPHOPODA	*Siphonodentalium* sp.	7000–6900	Sunda	Gal.	1951	BRUUN, 1953
CEPHALOPODA (OCTOP.)	—	8100	Kur.–Kamt.	Vit.	1949	ZENK. *et al.*, 1955
CRINOIDEA	—	9715–9735	Idzu–Bonin	Vit.	1955	Exhibit., 1958
	Bathycrinus australis (A. H. Clark)	8210–8300	Kermadec	Gal.	1952	Gislén, 1956
ASTEROIDEA	—	7585–7615	Mariana	Vit.	1955	Exhibit., 1958
	Eremicaster vicinus (Ludwig)	7000–6900	Sunda	Gal.	1951	MADSEN (*in litt.*)
OPHIUROIDEA	—	8006	New Britain	Vit.	1957	Exhibit., 1958
	Ophiura loveni (Lyman)	6660–6770	Kermadec	Gal.	1952	MADSEN, 1956
ECHINOIDEA	—	7290–7250	Banda	Gal.	1951	MADSEN, 1956
	Pourtalesia sp. ? *aurorae* Koehler	7290–7250	Banda	Gal.	1957	MADSEN, 1956
HOLOTHURIOIDEA	—	10,630–10,710	Mariana	Vit.	1958	Exhibit., 1958
	Myriotrochus bruuni Hansen	10,120–10,210	Philippine	Gal.	1951	HANSEN, 1956
POGONOPHORA	—	9700–9950	Kur.–Kamt.	Vit.	1953	IVANOV, 1957
	Zenkevitchiana longissima Ivanov	9700–9950	Kur.–Kamt.	Vit.	1953	IVANOV, 1957
ENTEROPNEUSTA	—	8100	Kur.–Kamt.	Vit.	1949	ZENK. *et al.*, 1955
TUNICATA	—	7210-7230	Kur.–Kamt.	Vit.	1953	ZENK. *et al.*, 1955
	Cnemidocarpa bythis (Herdman)	6960–7000	Kermadec	Gal.	1952	MILLAR, 1958
PISCES	—	7587	Japan	Vit.	1957	Exhibit., 1958
	Careproctus amblystomopsis Andriashev	7587	Japan	Vit.	1957	Exhibit., 1958

Vol. 6 1959

APPENDIX 3

List of trawlings and bottom samplings at depths exceeding 6000 m *with information about the following items, according to all records available : Depth, bottom temperature (in °Celcius), position, research vessel, station number, date, gear (see abbreviations below), number of species, approximate number of specimens and details about the fauna (owing to considerations of space the latter is indicated only by means of ciphers which refer to the ones given at the extreme left in Appendix* 1 *with the total list of the trench animals ; ciphers in italics refer to identified species). After the name of each trench is given its greatest depth recorded*).*

Abbreviations : HOT : Herring Otter Trawl ; ST : Sledge (or Agassiz or Sigsbee) Trawl ; D : Dredge ; PG : Petersen Grab (bottom sampler 0·2 or 0·25 m²).

PACIFIC OCEAN

ALEUTIAN TRENCH (7679 m)

6410 m *Vitjaz* Stn. 3340 1955 ST
Holothur. : *259*

7246 m **1·62 °C** 52° 25·8′N, 170° 53·9′E *Vitjaz* Stn. 3357 1955 ST + PG

Actiniaria : 50
Polychaeta : *82*
Echiuroid. : *114–115–117*
Isopoda : *156–159*
Lamellibr. : *211*

KURILE-KAMTCHATKA TRENCH (10,382 m)

c. 6000 m *Vitjaz* Stn. 3168 1955 PG

6860 m 48° 25·2′N, 156° 34·2′E *Vitjaz* Stn. 2144 (1–2.6.1953) ST

No. of species : **45 + 8**
Foramin. : *3–6–8–12–17–19–21–27*
Porifera : 32–34
Hydrozoa : 38
Scyphozoa : 40
Anthozoa : 44–56
No. of specimens : **c. 425**
Polychaeta : 64–110
Echiuroid. : *115*
Sipuncul. : 119–*120–121*
Cirripedia : *126*
Isopoda : *154*
Pycnogon : *184*
Gastropoda : 192–193
Lamellibr. : 208–217
Crinoidea : 231
Asteroidea : 235–*236*–241
Ophiuroid. : 244
Holothur. : 251–276

c. 7000 m *Vitjaz* Stn. 3257 1955 PG

7210–7230 m **1·78 °C** 49° 29·3′N, 158° 41·0′E *Vitjaz* Stn. 2208 22.6.1953 ST

No. of species : **42 + 13**
Foramin. : *3–5–8–10–13–15–16–17–19–20–21–22–27*
Hydrozoa : 35
Anthozoa : 51–56
Nemertini : 61
Polychaeta : 64–69–*84–89–90–*92–104–110
No. of specimens : **c. 690**
Echiuroid. : *118*
Priapul. : 123
Isopoda : *137–148–156–158*
Amphipoda : 161
Mysidacea : *182*
Gastropoda : 187
Lamellibr. : 209–212–217
Asteroidea : 241
Ophiuroid. : 244
Holothur. : 251
Tunicata : 285
Pisces : *288*

8100 m 43° 39′N, 152° 03′E *Vitjaz* Stn. 162 10.10.1949 ST

No. of species : **17 + 11**
Foramin. : 4–7–*8*–9–11–*21–*23–*24*–28–*29*
Porifera : 32
Actiniaria : 44
Polychaeta : *72*–81–*85–88*–102–*111*
No. of specimens : **c. 200**
Echiuroid. : *116*
Isopoda : *156*
Amphipoda : 161
Lamellibr. : 204
Cephalop. : 228
Holothur. : *259*–273
Pogonoph. : *277–280*
Enteropn. : 283

*After ZENKEVICH, 1958b.

8330–8430 m 46° 31′N, 154° 22′E *Vitjaz* Stn. 2120 26.5.1953 ST
No. of species : **20** No. of specimens : **c. 144**
Polychaeta : 69–*90*–110 Amphipoda : 161 Holothur. : 251–259
Echiuroid. : *115* Gastropoda : 187 Pogonoph. : *282*
Isopoda : *142*–145–*148*–*160* Lamellibr. : 198–208

8610–8660 m 1·99°C 45° 41·4′N, 153° 23·8′E *Vitjaz* Stn. 2216 26.6.1953 ST
No. of species : **9** No. of specimens : **c. 290**
Porifera : 32 Polychaeta : 64 Lamellibr. : 198
Hexacorr. : 43 Echiuroid. : *118* Holothur. : *259*–260
Octocorr. : 59 Amphipoda : 161

8632–8779 m 44° 08·5′N, 150° 22′E *Vitjaz* Stn. 3176 6.10.1955 ST
Anthozoa : *58* Echiuroid. : *118* Holothur. : *259*

9000–9050 m 2·08°C 44° 07·8′N, 150° 32·0′E *Vitjaz* Stn. 2217 29.6.1953 ST
No. of species : **18 + 5** No. of specimens : **c. 3200**
Foramin. : 14–18–*21*–*25*–*26* Echiuroid. : *118* Crinoidea : 233
Actiniaria : 44 Gastropoda : 187 Holothur. : 251–*259*
Polychaeta : 64–*88* Lamellibr. : 198–*207* Pogonoph. : *281*–*282*

9700–9950 m 2·1°C 43° 47·6′N, 149° 54·8′E *Vitjaz* Stn. 2218 1.7.1953 ST
No. of species : **4** No. of specimens : **c. 250**
Polychaeta : *88* Echiuroid. : *118* Holothur. : 251–*259*

JAPAN TRENCH (8597 m)

6156–6207 m 38° 10·9′N, 143° 56·2′E *Vitjaz* Stn. 3214 25.10.1955 ST
Actiniaria : 52 Isopoda : *153* Pogonoph. : *279*
Echiuroid. : *116* Lamellibr. : *201*–*219* Tunicata : 285
Sipuncul. : *120*–*121* Holothur. : *259* Pisces : *288*
Cirripedia : *126*

6475–6570 m 41° 22·9′N, 145° 15·4′E *Vitjaz* Stn. 3457 21.9.1955 ST
Actiniaria : 52 Echiuroid. : *116* Isopoda : 143–*153*–*156*

7190–7280 m 38° 02′N, 143° 57·4′E *Vitjaz* Stn. 3227 2.5.1955 ST
Isopoda : *156* Holothur. : *259*

7587 m 33° 02·6′N, 144° 58·7′E *Vitjaz* Stn. 3571 1957 ST
Echiuroid. : *115* Lamellibr. : *220* Pisces : *288*–289
Priapul. : 123 Holothur. : *259*

N.W. PACIFIC

6096 m 33° 18′N, 149° 46′E *Vitjaz* Stn. 3232 6.5.1955 ST
Cirripedia : *126* Lamellibr. : *221*

6272–6282 m 48° 15′N, 169° 39′E *Vitjaz* Stn. 3363 10.6.1955 ST
Sipuncul. : *120* Asteroidea : 239–242 Tunicata : 285
Lamellibr. : *219*

IDZU - BONIN TRENCH (9764 m)

7305–7315 m 30° 34′N, 142° 41′E *Vitjaz* Stn. 3491 7.10.1955 ST
Isopoda : *160*

9600 m (8540 m ?) *Vitjaz* Stn. 3514 1955 ST
Holothur. : *259* Pogonoph. : *278*

9715–9735 m *Vitjaz* Stn. 3494 1955 ST
Echiuroid. : *118* Crinoidea : 230 Holothur. : *259*

Vol 6 1959

RIU-KIU TRENCH (7507 m)

6810 m (?) **1·84°C** — 27° 30′N, 130° 35′E — *Vitjaz* Stn. 3528 28.10.1955 ST
Lamellibr. : *218*

MARIANA TRENCH (10990 m)

7585–7615 m — *Vitjaz* Stn. 3689 1955 ST
Asteroidea : 234

10630–10710 m — *Vitjaz* Stn. 4004 1958 ST
No. of species : **4** (? 6) — No. of specimens : **12**
Actiniaria : 53 — Polychaeta : 86 — Holothur. : 253–263

PHILIPPINE TRENCH (10497 m)

9820–10,000 m **2·6°C** — 10° 20′N, 126° 41′E — *Galathea* Stn. 435 7.8.1951 ST
No. of species : **5** — No. of specimens : **8**
Actiniaria : *48* — Amphipoda : *174* — Holothur. : *274*
Isopoda : *150* — Lamellibr. : 203

10,020–10,120 m **2·6°C** — 9° 49′N, 126° 44′E — *Galathea* Stn. 429 2.8.1951 D
No. of species : **1** — No. of specimens : **3**
Amphipoda : *169*

10120 m **2·6°C** — 10° 28′N, 126° 39′E — *Galathea* Stn. 424 27.7.1951 PG
No. of species : **1** — No. of specimens : **1**
Holothur. : *266*

10190 m **2·6°C** — 10° 13′N, 126° 43′E — *Galathea* Stn. 418 21.7.1951 ST
No. of species : **5** — No. of specimens : **133**
Actiniaria : *48* — Amphipoda : *169* — Holothur. : *266*
Polychaeta : *83* — Lamellibr. : 203

10150–10210 m **2·6°C** — 10° 19′N, 126° 39′E — *Galathea* Stn. 419 22.7.1951 ST
No. of species : **4** — No. of specimens : **24**
Actiniaria : *48* — Echiuroid. : 112 — Holothur. : *266*
Polychaeta : *83*

BANDA TRENCH (7298 m)

6490–6650 m **3·5°C** — 5° 18′S, 131° 18′E — *Galathea* Stn. 497 23.9.1951 HOT
No. of species : **10** — No. of specimens : **c. 70**
Scyphozoa : *41* — Amphipoda : *172* — Holothur. : *254–255–261–265–268*
Polychaeta : *99–109* — Pychnogon. : *183*

6580 m **3·5°C** — 5° 21′S, 131° 17′E — *Galathea* Stn. 499 24.9.1951 PG
No. of species : **8** — No. of specimens : **12**
Polychaeta : *73–99–108–109* — Amphipoda : *167* — Holothur. : *255*
Isopoda : *149* — Lamellibr. : 202

7270 m **3·6°C** — 5° 36′S, 131° 06′E — *Galathea* Stn. 496 23.9.1951 PG
No. of species : **3** — No. of specimens : **5**
Isopoda : *151* — Amphipoda : *167* — Lamellibr. : 202

7240–7290 m **3·6°C** — 5° 36′S, 131° 01′E — *Galathea* Stn. 494 22.9.1951 ST
No. of species : **3** — No. of specimens : **4**
Polychaeta : *78–100* — Holothur. : *268*

7290–7250 m **3·6°C** — 5° 26′S, 130° 58′E — *Galathea* Stn. 495 22.9.1951 HOT
No. of species : **14** — No. of specimens : **c. 100**
Actiniaria : *47* — Amphipoda : *164* — Echinoidea : *250*
Polychaeta : *67–82–83–96–100–109* — Lamellibr. : 222–225 — Holothur. : *254–261–268*

NEW BRITAIN TRENCH (9140 m)

6920 m *Vitjaz* Stn. 3655 1957 **ST**

7657 m *Vitjaz* Stn. 3663 1957 **ST**
Ostracoda : 125 Loricata : 185 Ophiuroid : 245 (8006 m ?)
Cumacea : 129 (8006 m ?)

8830–8780 m **2·7 C** 5° 59′S, 153° 28′E *Galathea* Stn. 521 16.10.1951 **ST**
No. of species : **2** No. of specimens : **76**
Holothur. : *257–264*

8940 m **2·7 C** 6° 31′S, 153° 58′E *Galathea* Stn. 517 13.10.1951 **ST**
No. of species : **4** No. of specimens : **19**
Polychaeta : 66 (tube) Holothur. : *257–264–266*

8980–9043 m *Vitjaz* Stn. 3663–A 1957 **ST**
Polychaeta : 66–68–76–95 Gastropoda : 189 Holothur. : *259–262*
Isopoda : 136

NEW HEBRIDES TRENCH (7570 m)

6680–6830 m *Vitjaz* 1958 **ST**

TONGA TRENCH (10840 m)

7345 m *Vitjaz* Stn. 3818 1957 **ST**

c. 9700 m *Vitjaz* Stn. 3823 25.12.1957 **ST**
Polychaeta : 65

10687–10415 m *Vitjaz* Stn. 3823–A 28.12.1957 **ST**
No. of species : **17 + 4** No. of specimens : **c. 100**
Foramin. : 2 Isopoda : 134 Lamellibr. : 199
Porifera : 33 Amphipoda : 162–175–176 Holothur. : 267
Nematodes : 62 Gastropoda : 189
Polychaeta : 65–75–101–103

KERMADEC TRENCH (10030 m)

6180 m **1·2 C** 32° 09′S, 176° 35′W *Galathea* Stn. 653 17.2.1952 **HOT**
No. of species : **10** No. of specimens : **63**
Scyphozoa : *41* Polychaeta : 93–*94* Holothur. : *269–275*
Actiniaria : *48* Amphipoda : *174* Tunicata : *281*
Octocorr. : 59 Asteroidea : *237*

6620–6730 m **1·3 C** 32° 20′S, 176° 54′W *Galathea* Stn. 650 15.2.1952 **ST**
No. of species : **24** No. of specimens : **c. 425**
Porifera : 30 Polychaeta : 79–*82–91*–93– Gastropoda : 196
Scyphozoa : *41* *94–97*–105 Lamellibr. : 205–216
Actiniaria : *49* Cirripedia : 127 Asteroidea : *236*
Octocorr. : 59 Isopoda : *152–157* Ophiuroid. : 246
Nematodes : 63 Amphipoda : *177* Holothur. : *256–269–275*

6660–6770 m **1·3 C** 35° 51′S, 178° 31′W *Galathea* Stn. 658 20.2.1952 **ST**
No. of species : **32** No. of specimens : **c. 1190**
Porifera : 30 Cirripedia : 127 Lamellibr. : 214–224
Hydrozoa : *37* Tanaidacea : *131* Asteroidea : *238*
Actiniaria : 45–*49–54* Isopoda : *144–147–152–155* Ophiuroid. : *249*
Polychaeta : 71–*82–83–91–* Amphipoda : *170–181* Holothur. : *269–272–275*
94–97–106 Gastropoda : 191–194–197 Pisces : 290
Echinoid. : 113

Vol 6 1959

6960–7000 m **1·3 C** 32° 10′S, 177° 14′W *Galathea* Stn. 651 16.2.1952 **HOT**
No. of species : **31**
Porifera : 30
Scyphozoa : *41*
Actiniaria : *49*
Polychaeta : *80–82–83–94–97*
Echiuroid. : 113
No. of specimens : **c. 235**
Cirripedia : 127
Tanaid. : *132–133*
Isopoda : *139–140–141–144–146–147–152*
Amphipoda : *163–165–177–179–180*
Lamellibr. : 210–215–216
Holothur. : *269–272*
Tunicata : *284*

7640 m **1·4 C** 35° 20′S, 178° 55′W *Galathea* Stn. 656 20.2.1952 **ST**
No. of species : **1**
Amphipoda : *168*
No. of specimens : **1**

8210–8300 m **1·5 C** 35° 16′S, 178° 40′W *Galathea* Stn. 649 14.2.1952 **ST**
No. of species : **20**
Hydrozoa : *39*
Actiniaria : *46–49*
Polychaeta : 79–*80–83*–93–98
Echiuroid. : 113
No. of specimens : **c. 2100**
Tanaidacea : *133*
Amphipoda : *173–177*
Gastropoda : 195
Lamellibr. : 205–210–226
Bryozoa : 229
Crinoidea : *232*
Holothur. : *256–269*

8928–9174 m *Vitjaz* Stn. 3827 1958 **ST**
Tanaidacea : 130

9995–10002 m *Vitjaz* Stn. 3831 1958 **ST**
No. of species : **11 + 5**
Foramin. : 1
Scyphozoa : 42
Nematodes : 63
No. of specimens : **177**
Polychaeta : 74
Copepoda : 124
Isopoda : 135
Gastropoda : 190
Lamellibr. : 200–223
Holothur. : 252

Vol. 6 959-60

INDIAN OCEAN

SUNDA TRENCH (7317 m)

6730 m **1·4 C** 10° 02′S, 107° 52′E *Galathea* Stn. 462 3.9.1951 **PG**
No. of species : 3
Porifera : 31
No. of specimens : 3
Polychaeta : *77–80*

7000–6900 m **1·4 C** 10° 20′S, 109° 55′E *Galathea* Stn. 465 5.9.1951 **D**
No. of species : 7
Hydrozoa : *36*
Anthozoa : 57
Polychaeta : 70
No. of specimens : **28**
Lamellibr. : 206
Scaphopoda : 227
Asteroidea : *236*
Holothur. : *258*

7160 m **1·4 C** 10° 21′S, 110° 12′E *Galathea* Stn. 466 6.9.1951 **HOT**
No. of species : 7
Actiniaria : *55*
Polychaeta : *82*
No. of specimens : **c. 3170**
Cumacea : 129
Amphipoda : *178*
Holothur. : *258–271*
Pisces : 287

ATLANTIC OCEAN

PUERTO RICO TRENCH (9218 m)

7625–7900 m 19° 49′N, 65° 01′W–19° 42′N, 64° 44′W *Albatross* Stn. 370 17.8.1948 **SOT**
No. of species : 7
Polychaeta : 87–107
Isopoda : *138*
No. of specimens : **c. 35**
Amphipoda : *163–166–171*
Holothur. : 270

CAPE VERDE BASIN

6035 m 12° 07′30″N, 33° 32′45″W *Princesse–Alice* 6.8.1901 **ST**
No. of species : **4**
Asteroidea : *240*
No. of specimens : **6**
Ophiuroid. : *247–248*
Pisces : *286*

APPENDIX 4

The hadopelagic fauna recorded from a haul with closing net from 6000–8500 m *depth in the Kurile-Kamtchatka Trench.*

	No. of specimens	*Recorded by*
CRUSTACEA		
COPEPODA		
Aetidius sp. ?	1	BRODSKY, 1955
Calanus tonsus Brady, 1883	1	,,
Heterorhabdus compactus (Sars, 1900)	2	,,
Lucicutia ushakovi Brodsky, 1955	1	,,
Metridia okhotensis Brodsky, 1950	1	,,
Metridia similis Brodsky, var. *abyssalis* Brodsky, 1955	37	,,
Mimocalanus distinctocephalus Brodsky, 1950	1	,,
Parascaphocalanus zenkevitchi Brodsky, 1955	97	,,
Pareuchaeta sp.	2	,,
Pseudeuchaeta sp.	1	,,
Pseudochirella gurjanovae Brodsky, 1955	13	,,
Scaphocalanus bogorovi Brodsky, 1955	1	,,
Scaphocalanus sp.	1	,,
Scolecithrix birshteini f. *major* Brodsky, 1955	1	,,
Scolecithrix birshteini f. *minor* Brodsky, 1955	1	,,
Scolecithricidae gen. I	1	,,
Scolecithricidae gen. II	1	,,
Spinocalanus similis Brodsky, var. *profundalis* Brodsky, 1955	64	,,
Xanthocalanus pavlovskii Brodsky, 1955	2	,,
Zenkevitchiella abyssalis Brodsky, 1955	1	,,
OSTRACODA (4 spp.)		BIRSTEIN *et al.*, 1954
AMPHIPODA		
Andaniexis subabyssi Bst. and Vinogr., 1955	2	BIRSTEIN and VINOGRADOV, 1955
Halice quarta Bst. and Vinogr., 1955	4	,,
Hirondellea gigas Bst. and Vinogr., 1955	1	,, (DAHL, 1959)
Hyperiopsis laticarpa Bst. and Vinogr., 1955	2	,,
Vitjaziana gurjanovae Bst. and Vinogr., 1955	—	,,

Vol 6 1959

The relationship of the distribution of the planktonic worm, *Poeobius meseres* Heath, to the water masses of the North Pacific

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Abstract—The distribution of a planktonic worm, *Poeobius meseres* Heath, was determined from an examination of over 1800 quantitative plankton tows taken in the North and South Pacific. This distribution is compared with the distribution of water masses, as defined by temperature-salinity curves. Since the water mass concept involves a three dimensional unit of the ocean, its use in describing the environment of an animal whose distribution is also three dimensional, is preferable to the method of comparing plankton distributions with horizontal isotherms or isohalines.

The distribution of *Poeobius* coincides, for the most part, with that of the Subarctic water mass and the transition region of Subarctic water, the California Current. However, a few specimens were found in the eastern tropical Pacific. A satisfactory explanation for its restricted presence in this latter area is not possible at this time, but there is some evidence to indicate that this southern segment of the population is not endemic but has been carried in from the north. If this is true, then the occurrence of *Poeobius* here must be accounted for in considering the sources of the Intermediate water of the area.

Vol.
6
959-60

INTRODUCTION

The environment of most oceanic, zooplanktonic species is generally described in terms of the temperature and salinity ranges to which the animals seem limited. But because zooplankton is distributed three-dimensionally it is frequently difficult to relate the overall horizontal distributions of species to the horizontal distributions of temperature or salinity limits. However, with the development of the water mass concept (Helland-Hansen 1916, Sverdrup *et al.*, 1942) it has become possible to relate the distributions of species to that of water masses. It is the purpose of this paper to present the distribution of a pelagic organism, *Poeobius meseres* Heath (1930), and to compare it with the distribution of a water mass.

Poeobius meseres, a small, transparent, planktonic animal ranging in length up to 27 mm, was originally described by Heath (1930) from Monterey Bay, California. He obtained specimens from a plankton sample taken from a depth of approximately 350 m. Heath suggested that *Poeobius* was " a connecting link between the Annelida and the Echiuroida." Although Remane (see Pickford, 1947) considered it to be an echiuroid, Fisher (1946) excluded it from that Phylum. Pickford (1947), after an intensive study of the anatomy and histology of *Poeobius*, came to the conclusion that it was an aberrant Polychaete. Both Heath and Pickford were concerned primarily wth the phylogenetic status of this strange worm, and virtually nothing is known of its ecology and life history. The only information available on the distribution of *Poeobius* is the statement by Pickford (1947) that " Pelagic in habit, it has been taken from moderate depths off the coasts of California and in Alaskan waters," the report by Bogorov (1955) of its occurrence in the North-west Pacific (position not given) and Hartman's (1955) report that " *Poeobius meseres* is known only from the northern Pacific."

During the analysis of over 1800 plankton samples taken in the Pacific by the Scripps Institution of Oceanography, the occurrence of *Poecobius* was recorded. These observations were then compared with hydrographic data which had been taken at the same time and place. This comparison revealed that the distribution of *Poecobius* generally coincided with the distribution of the Subarctic Water Mass.

METHODS

The plankton samples used in this study came from several different cruises and expeditions (Fig. 1, Table 1). The routine sampling done by the California Co-operative Oceanic Fisheries Investigations (C.C.O.F.I.), served as a model for all of these cruises

Table 1. *Plankton samples used in this study*

Expeditions Scripps Institution of Oceanography	*Date*	*Number of tows*	*Depth of tows* (m)
Mid-Pacific	1950	6 12	0-100 0-1000 or greater
Northern Holiday	1951	55 21 trawls (not used in this study)	0-140
Shellback	1952	203 15 trawls	0-300 0-1000 or greater
Capricorn	1953	35	0-140; 0-400; 0-1000
Trans-Pacific	1953	143 135 38 12 16 4 trawls	0-150 150-300 300-450 450-600 0-1000 0-2000 or greater
Troll	1955	85	0-300
EQUAPAC	1956	89	0-280
PAS	1954	14 12 14 13 20 10 12	0-25 25-50 50-75 75-100 100-300 300-500 500-700
CCOFI	1950 to 1954	532 (used in this study)	0-140
CCOFI	1950	8	0-400
Nor-Pac	1955	162 30 20	0-140 140-280 0-700
Downwind	1957	162 30 20 12	0-140 140-280 0-560 0-700 or greater
Tage (Hopkins Marine Sta.)	1951-1952	8	0-700 or greater

Vol 6 1959

and expeditions (AHLSTROM, 1954). The vertical distribution of *Poeobius* was determined, for the most part, from the results of hauls made with an opening-closing net, one metre in diameter, adapted from that described by LEAVITT (1935, 1938). This net may be towed obliquely and opened and closed at any desired depth. Occasionally a number of these nets were used in series along a cable, thus sampling several strata simultaneously (Fig. 2).

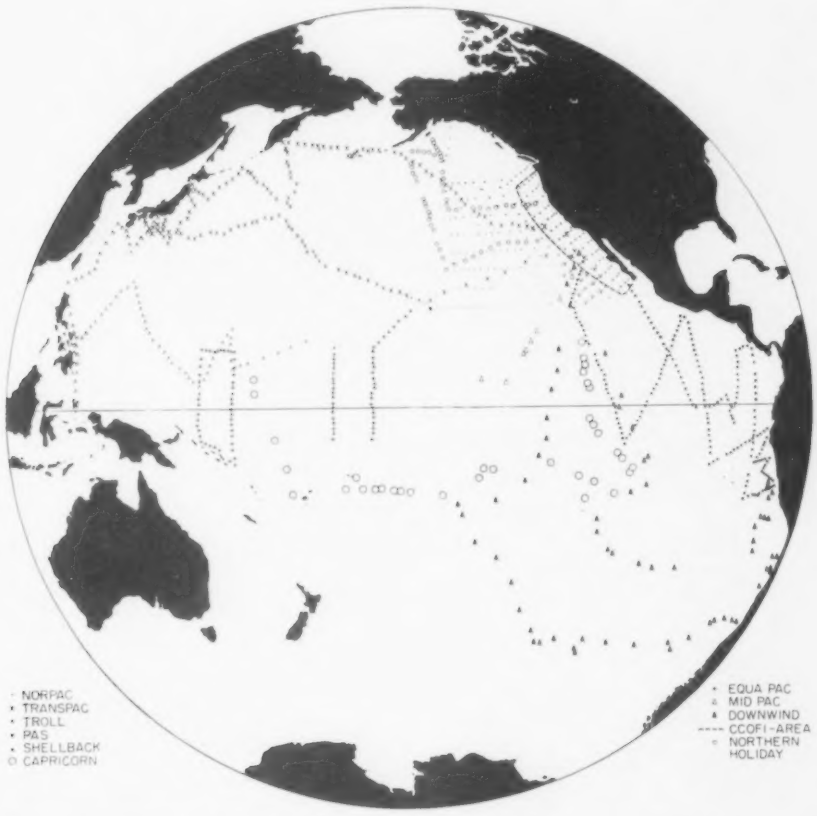

Fig. 1. Scripps expeditions in the North Pacific. Quantitative plankton hauls were taken on every station shown.

DISTRIBUTION

(1) *Geographic distribution*

The greatest density of *Poeobius* was found in the North Pacific in the region of the Oyashio and Subarctic current. Although DALES (1957) did not find it in the surface waters (0-70 m) of the California Current, it does occur at intermediate depths (400 m or greater) in this current and also in the eastern-most part of the tropical Pacific but in greatly decreased numbers (Fig. 3). It is probably present in the Gulf of Alaska but few samples have been taken from that region. *Poeobius* did not occur in any of the 298 plankton hauls taken from the upper 700 m of the

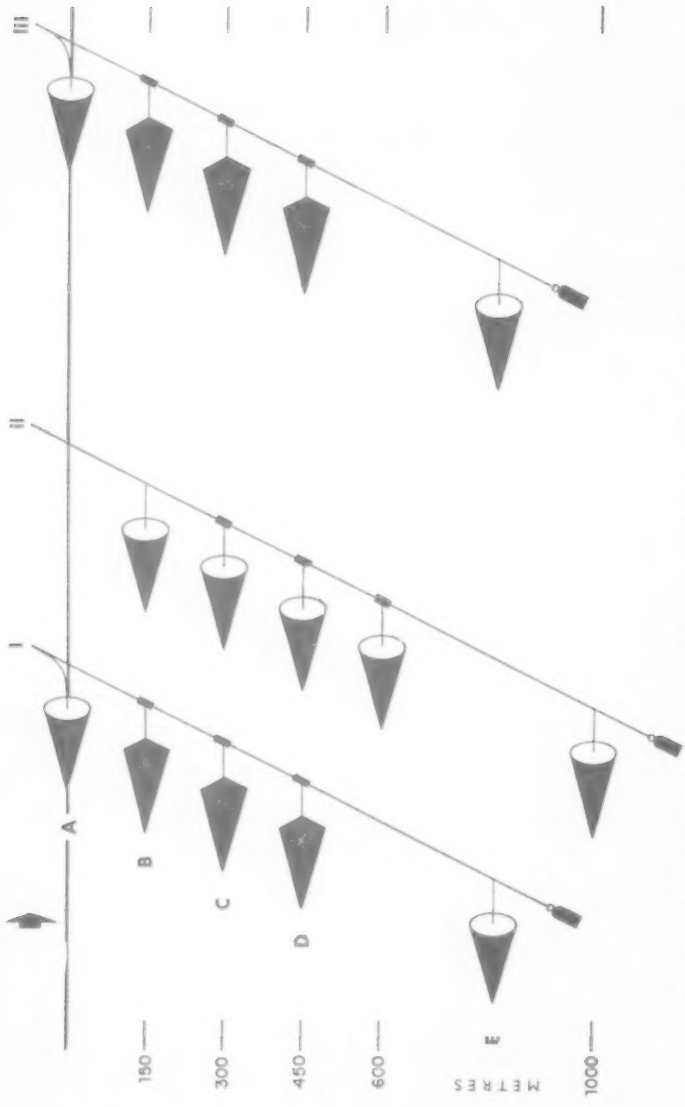

Fig. 2. Diagram illustrating an opening-closing series as taken on the Trans-Pacific Expedition. Nets A and E were standard, quantitative nets. Nets B, C, and D were opening-closing nets. A 600 lb weight was attached to the bottom of the wire. At I, just before net A entered the water a messenger was sent down to open nets B, C, and D, then the entire series was lowered 150 m (point II). Between points II and III the series was raised 75 m and towed at that depth for approximately 20 minutes. At point III the series was raised another 75 m, net A was removed from the water and a second messenger was sent down to close nets B, C, and D. The entire series was then removed from the water. Nets A, B, C, and D each fished a stratum 150 m thick while net E fished from the surface to depths of 1000 m or more. The ship maintained a speed of 1·1·5 knots during these tows. The entire procedure took approximately 2 hr.

Vol 6 1959

Eastern or Western North Pacific Central waters, nor did JOHNSON (1956) record it as being present in the Beaufort or Chukchi seas. It apparently does not exist in any other oceans of the world.

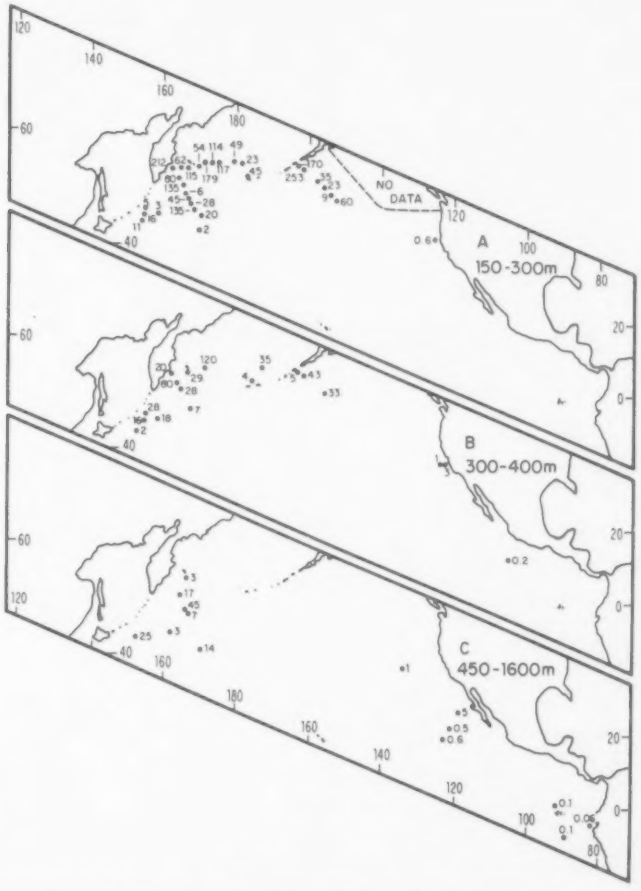

Fig. 3. The geographical distribution of *Poeobius*. The numbers of organisms per 1000 m^3 of water filtered by the net are shown for three strata : A, 150-300 m ; B, 300-450 ; C, 450-1600 m. By increasing the depth of tow, the range is extended to the south in the eastern Pacific, and the abundance of *Poeobius* decreases.

(II) *Vertical distribution*

The opening-closing net tows taken on the Trans-Pacific Expedition show that in its area of greatest abundance in the Oyashio and Subarctic current, *Poeobius* is most commonly found at depths ranging from 150 to 300 m (Table 2 and Fig. 6). Although no systematic use of the opening-closing net has been made in the California Current, the area has been intensively sampled with open, quantitative, one-metre nets towed at various depths (U.S. Fish and Wildlife Service, 1954). In the northern part of this region (Tables 1 and 3), *Poeobius* was found only in those tows that went to a depth of 400 m or more ; in the southern part, which was covered by the Pelagic

Area Studies (P.A.S.) survey (Fig. 1), *Poeobius* was found only at depths of 700 m or more. South of 20° N in the eastern tropical Pacific* it was found, with one

Table 2. Selected Trans-Pacific Expedition opening-closing series, showing depth and temperature range of Poeobius

Station	*Depth of tow* (m)	*Numbers per* 1000 m^3	*Temperature range* (°C)
25 A	0-150	0	3·5-11·2
B	150-300	170	4·0-4·3
C	300-450	5	3·7-4·0
D	450-600	5	3·4-3·7
40 A	0-150	0	1·4-9·1
B	150-300	179	3·25-3·6
C	300-450	120	3·3-3·6
D	450-600	0	3·1-3·3
42 A	0-150	0	0·6-9·2
B	150-300	115	0·9-3·6
C	300-450	29	3·4-3·6
D	450-600	0	3·1-3·4
46 A	0-150	0	0·7-9·9
B	150-300	135	3·1-3·35
C	300-450	28	3·2-3·4
D	450-600	17	3·0-3·25
48 A	0-25	0	9·2-9·7
B	25-100	0	1·3-9·3
C	35-115	0	1·3-7·0
D	140-320	40	1·7-3 6
59 A	0-150	0	5·6-14·65
B	150-300	0	4·1-5·55
C	300-450	0	3·8-4·1
D	450-600	3	3·55-3·8

exception, only in trawls that reached depths of 1000 m or more (Fig. 3). The exception was the capture of two individuals at a depth of 300 m near Shellback Station 207 (Table 3). WOOSTER and CROMWELL (1958, p. 178) have pointed out that this area near the northern boundary of the Equatorial Countercurrent is characterized by very shallow thermoclines.

Relationship of Poeobius *distribution to water masses*

The water in the top 1000 m north of 45° N in the Pacific has been called the Subarctic Water Mass (SVERDRUP *et al.*, 1942, p. 712). It is generally believed that this water originates mostly from the Oyashio with some slight mixing with water from the Kuroshio Extension. As this Subarctic water approaches the North American continent it splits, some moving northward to form the Gulf of Alaska Gyre, the rest moving southward as the California Current. The temperature-salinity relationship gradually changes as the water moves south and by the time the California Current begins to turn south-west at about 25° N latitude, becoming part of the North

* WOOSTER and CROMWELL (1958) have defined this area as " the region lying between the Tropic of Cancer (23° 27′ N) and the Tropic of Capricorn (23° 27′ S) and extending westward from the coast of Central and South America to 130° W.

Vol 6 1959

Equatorial Current (Reid *et al.*, 1958), mixing has considerably altered its properties. Ninety-seven per cent of all *Poeobius* caught were in hauls taken in the area of the

Table 3. *All* Poeobius *records excepting the Trans-Pacific Expedition, the Norpac Expedition, and those reported by* PICKFORD (1947)

Cruise station	*Position*	*Date*	*Depth of haul* (m)	*Numbers of organisms caught*	*Water filtered by net* (m^3)
CCOFI 5201 137·40	25° 00′N 113° 25·5′W	1-52	0-147	1	444
CCOFI 5004 47·55	40° 04′N 124° 55′W	4-50	0-421	1	1668
CCOFI 5004 60·70	37° 17′N 124° 21′W	4-50	0-423	3	1945
N. H.	53° 34·5′N 155° 00′W	8-51	0-1200	several	—
Tage 248-B	Monterey Bay	9-52	0-740	25	4721
Tage 242-B	Monterey Bay	3-52	0-910	12	4904
Tage 144-B	Monterey Bay	12-51	0-980	3	4941
PAS 120·90	26° 12′N 118° 30′W	4-54	0-900	1	209
PAS 130·150	22° 30′N 121° 15′W	4-54	0-700	1	2160
PAS 140·180	19° 40′N 122° 15′W	4-54	0-700	2	3475
SB 186-187	1° 1·5′N 91° 45·7′W	8-52	0-938	2	—
SB 108-109	4° 04′S 82° 14′W	7-52	0-1640	1	—
SB 207	12° 13·5′N 106° 07′W	8-52	0-300	2	—
SB 101-102	06° 58′S 88° 35′W	7-52	0-1463	2	—

Subarctic Water Mass (Fig. 4). Temperature-salinity curves based on data taken at these stations where the depth of *Poeobius* capture was known, fell generally within the range of curves characteristic of this water mass (Fig. 5). Those few curves which fell outside this range were from stations on the edge of the *Poeobius* distribution, and even these curves indicate the presence of a certain amount of Subarctic water. Further evidence that *Poeobius* resides primarily in Subarctic Water is the finding, based on the results of the Trans-Pacific Expedition, that *Poeobius* disappeared abruptly from the plankton when the relatively sharp physical boundaries were crossed from Subarctic to Central waters (Figs. 6 and 7). On the other hand, in the areas where there is a gradual change of properties from the Subarctic Water to the California Current, the numbers of *Poeobius* decreased gradually (Fig. 3).

There is some reason to believe that *Poeobius* might occur in the Intermediate Water of the eastern and western north-central Pacific. UDA (1938), SVERDRUP *et al.* (1942, p. 716) and others have described a convergence zone at about 41° N ,between 150° and 160° E, where mixed Oyashio and Kuroshio water sinks and spreads out

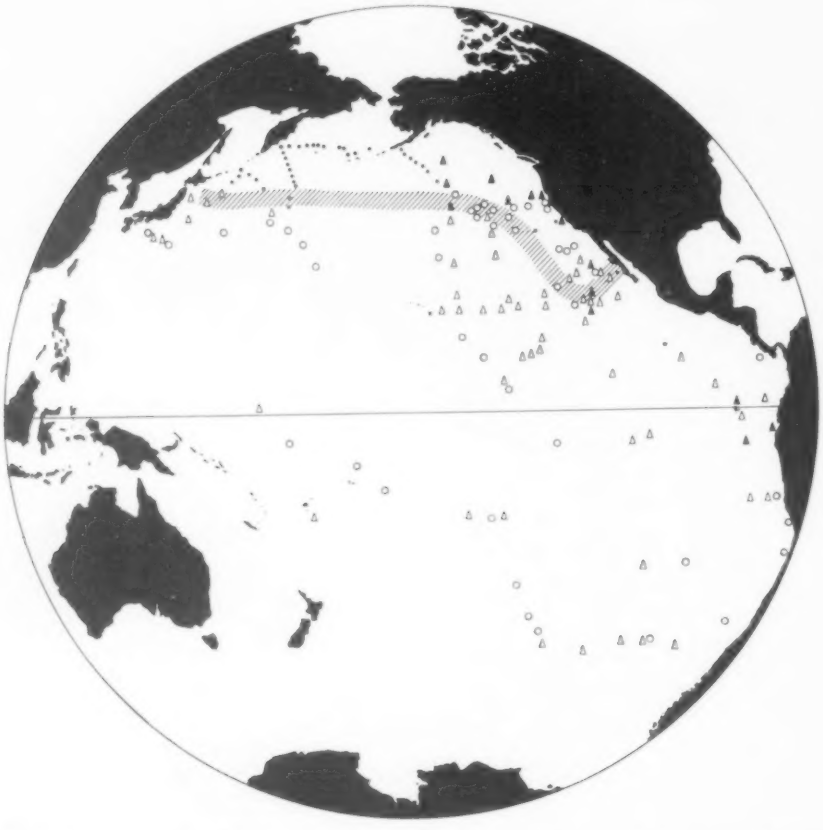

Fig. 4. All known records of *Poeobius*. Circles indicate net tows that reached a depth of 400 m or more, triangles indicate tows that reached a depth of 700 m or more. Those circles and triangles shown in solid black indicate stations where *Poeobius* were captured. Open circles and triangles indicate that no *Poeobius* were found at that station. The hatched portion of the chart is the boundary area of Subarctic and Transition Region water (SVERDRUP *et al.*, 1942, p. 740).

in a clockwise circulation over most of the North Pacific at depths ranging from 200 to 900 m. Although the shallower strata (0-300 m) of this area have been fairly well sampled (Fig. 1), only 21, widely spaced tows reached a depth of 700 m or more in the eastern half of this region (Fig. 4). *Poeobius* was not found in any of these tows. If it is assumed that the animal would be randomly distributed at low densities, the failure to find it in these tows can be used to set an upper 95 per cent confidence level of 1 animal per 11,548 m^3 of water for its density in this region (FISHER and YATES, 1938, p.1). This may be compared with an average density of 1 animal per 17 m^3 in the Subarctic Water Mass.

Vol
6
1959

If there were no further records of *Poeobius*, its distribution could be simply related to that of the Subarctic Water Mass. However, seven individuals were found in the Intermediate Water of the eastern equatorial Pacific. There is little doubt that some mixing does occur between this water mass and the southern portion of the California Current, but this does not seem sufficient to account for the numbers of *Poeobius* caught here. The source of the Intermediate water of the eastern tropical

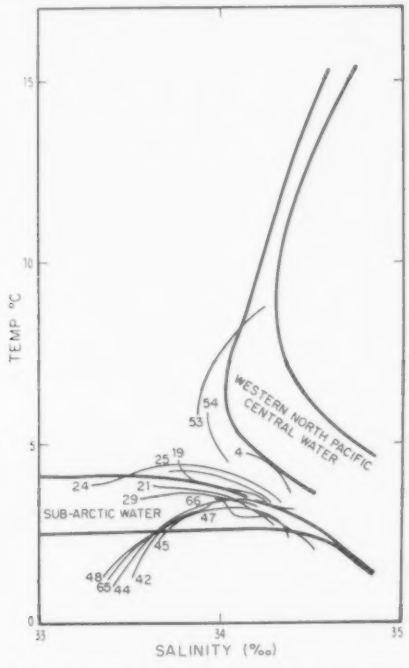

Fig. 5. Temperature-salinity curves from stations where *Poeobius* was taken on the Trans-Pacific Expedition, compared with SVERDRUP's Subarctic and Western North Pacific water. Station numbers are indicated on the fine lines. Only that segment of the curve within the depth range of *Poeobius* is plotted.

Pacific is not really known, but SVERDRUP suggests that it is " formed off the coast of South America by the gradual transformation of Subantarctic water " (SVERDRUP *et al.*, 1942, p. 706). If this is true, the presence of *Poeobius* might indicate that it is an antitropical species (HUBBS, 1952) as are several other pelagic organisms in the Pacific Ocean. However, plankton samples taken on the Scripps Downwind Expedition (Fig. 1) and reports of the *Dana* (Wesenburg-Lund, personal communication) and *Discovery* expeditions (MONRO, 1936) indicate that it is not present in either the Central Water or the Subantarctic water of the South Pacific, and TEBBLE (1958) did not find it in the Antarctic or Subantarctic waters of the South Atlantic. It seems therefore that this explanation cannot be used to account for the occurrence of *Poeobius* in the eastern tropical Pacific.

Another possible explanation for the presence of *Poeobius* in this area is that these seven individuals represent a second and hitherto undescribed species with different ecological requirements. There are however, only slight morphological bases for this

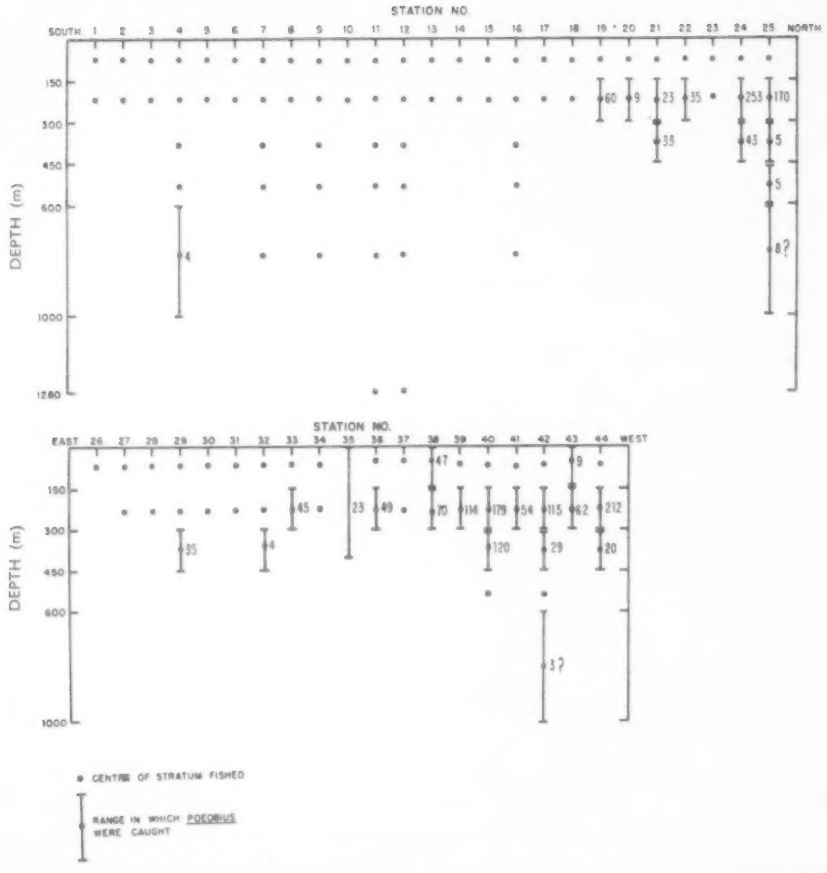

Fig. 6. Trans-Pacific sampling programme, Stations 1 to 44, showing the vertical distribution of *Poeobius*. The numbers to the right of the vertical bars indicate the numbers of organisms per 1000 m^3. The specimens from the deepest tows on Stations 25 and 42 might have come from the shallower depths.

Vol 6 1959

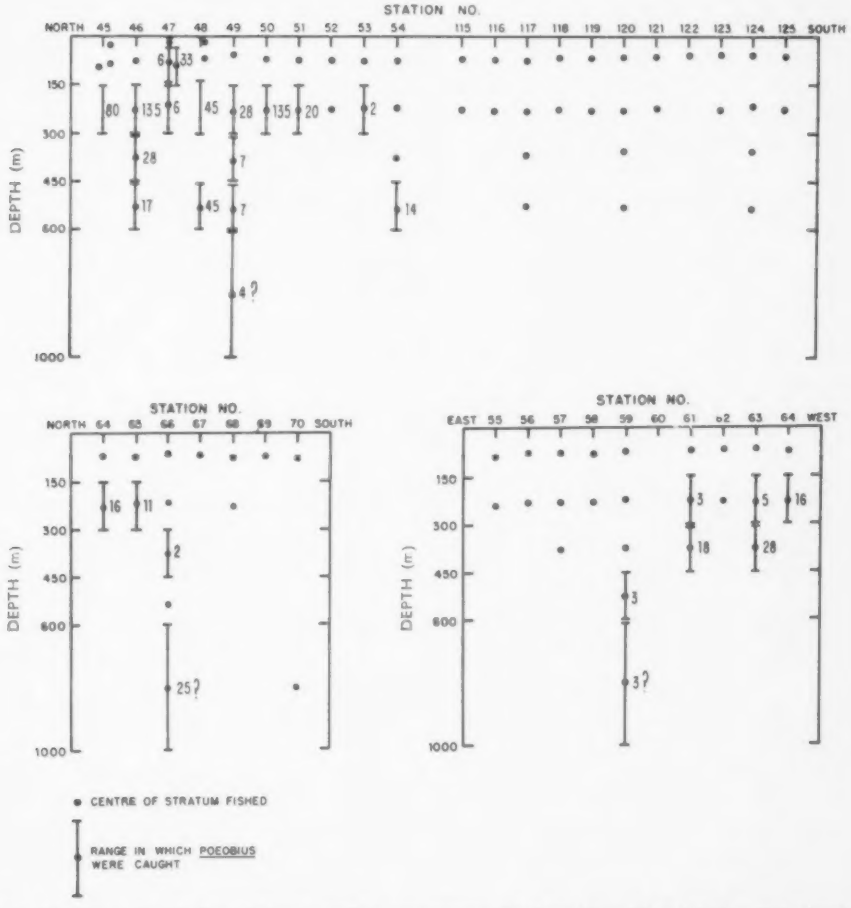

Fig. 7. The Trans-Pacific Expedition sampling programme, Stations 45 to 70 and 115 to 125, showing the vertical distribution of *Poeobius*. The specimens from the deepest tows on Stations 49 and 59 might have come from the shallower depths.

suggestion. All specimens from this area were rather larger than the average of those collected in pure Subarctic water, but despite the large size of these animals, their gonads were either missing or greatly reduced in size (Table 4). These differences do not necessarily indicate that they represent a new species for the resorption of gonads and other internal organs is a well known phenomenon in invertebrates living under unfavourable conditions. One might, therefore, expect to find such a condition in *Poeobius* that had been carried out of their normal environment. If this explanation is correct and the *Poeobius* in this area are sterile expatriates from the north, our ideas on the sources of the Intermediate water of the eastern tropical Pacific will have to be reconsidered.

DISCUSSION

Few attempts have been made to relate the entire distribution of a pelagic planktonic organism to the distribution of water mass*. RUSSELL (1935 and 1939), FRASER (1939 and 1952), PIERCE (1939) and others have shown that the distributions of chaetognaths and certain other organisms in the vicinity of the British Isles are related to the movements of waters from different sources and according to FRASER (1952) " it has now been established that the various species of Chaetognatha are associated each with a more or less specific hydrographical environment, making the group one of the foremost of the biological indicators." JOHNSON (1956) has shown that copepods also may be used as an aid in tracing the water movements in the Chukchi and Beaufort seas. It should be pointed out, however, that most of the species considered by these authors are more widely distributed than the limited areas they have investigated and that until all of the boundaries of their distributions are well known, it will be difficult to define in physical terms the regimen to which they are limited, or to show whether or not their overall ranges coincide with the concept of a physical water mass. HAFFNER's (1952) study on *Chauliodus* is perhaps more extensive, but because he does not have accurate data on depth of capture, it is not certain that his species are really limited to the depth ranges to which he asssign them. Further, his technique of using a single temperature-salinity measurement at the presumed depth of capture is probably not sufficient to identify a water mass, for SVERDRUP *et al.* (1942, p. 143) have pointed out that this may be done only in exceptional cases and that a T-S curve is usually required to define a water mass.

TEBBLE (1958) has reported that certain pelagic polychaetes in the South Atlantic and Antarctic oceans are limited to particular water masses, but at the time of this writing his detailed information has not been published. In a recent study on the chaetognaths of the Antarctic regions, DAVID (1958) has shown that the boundaries of the ranges of some specimens do, indeed, coincide with hydrographic boundaries and that some of these species are endemic to the Antarctic water.

While T-S relationships are used to identify water masses, this does not necessarily mean that the organisms limited to these water masses are limited by temperature and salinity alone. Water having characteristics that would fit certain points of the Subarctic T-S curve may be found in many parts of the world, and yet, so far as is known at present, *Poeobius meseres* occurs only in Pacific Subarctic water or in waters which may have Subarctic components. There are many possible explanations but until more is known of the biology of the animal, they can only be considered

* As defined by SVERDRUP *et al.* (1942, p. 143).

Vol 6 1959

Table 4. Size and gonad condition of Poeobius *from the Subarctic and Eastern Equatorial regions*

Station	*Zone*	*Size of specimens*	*Month*	*Condition of Gonads*
TP 40B	Subarctic	18 mm 20 mm 11 mm 15 mm	Aug.	ovary large 2 mm ovary large 2 mm ovary large 2 mm ovary large 2 mm
TP 36C	Subarctic	17 mm 9 mm 12 mm 10 mm 9 mm 16 mm 15 mm 10 mm 10 mm 10 mm 13 mm	Aug.	ovary large 2-3 mm ovary large 2 mm ovary large 2 mm ovary large 2 mm ovary large 2 mm specimen opaque ovary large 2 mm — ovary large 2 mm specimen damaged ovary large 3 mm
TP 41B	Subarctic	18 mm 16 mm 13 mm	Aug.	ovary large 2 mm ovary 1·5 mm ovary 1·75 mm
TP 47D	Oyashio	13 mm 12 mm 18 mm 19 mm	Sept.	ovary 1 mm ovary ruptured, eggs diffuse in body cavity ovary 1 mm ovary ruptured, eggs diffuse in body cavity
TP 25B	Subarctic	6 mm 13 mm 9 mm 14 mm 10 mm 6·5 mm	Aug.	body opaque body wall ruptured ovary 1·25 mm body opaque ovary 1·25 mm body opaque
TP 66E	Oyashio	10 mm 16 mm 10 mm 8 mm 12 mm 12 mm 15 mm 16 mm 12 mm	Sept.	body opaque ovary large 2·5 mm ovary large 2·5 mm ovary 1·5 mm ovary ruptured ovary large 3 mm ovary large 2 mm ovary large 2·5 mm entire posterior half is ovary
SB 101-2	Eastern Tropical Pacific	21 mm 20 mm	Aug.	viscera degenerate, gonads missing viscera degenerate, gonads missing
SB 108-9	E.T.P.	10 mm	July	much of viscera missing, gonads missing
SB 207	E.T.P.	10 mm 15 mm	Aug.	viscera partially gone, gonads missing ovary present 1·5 mm
SB 186-7	E.T.P.	10 mm 10 mm	July	gonads missing gonads missing

theoretical. One factor which may be of importance is the higher productivity and standing crop of the Subarctic, California Current, and eastern tropical Pacific areas (HOLMES, 1958) compared with that of the Central and Equatorial Pacific water masses (STEEMANN-NIELSEN, 1957). Thus the southward extension of the range of *Poeobius* from the Subarctic water mass to the California Current to the eastern tropical Pacific may have a simple relationship to the availability of food.

From the foregoing discussion it appears that water masses, as defined by hydrographers, may also exist as biological entities. The water mass concept, which defines a three-dimensional unit of the ocean, thus provides us with an opportunity to compare plankton distributions with that of circumscribed volumes of water, each of which has a unique set of physical properties.

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Vol
6
1959

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Vol.
6
959-60

On the abyssal circulation of the world ocean—I. Stationary planetary flow patterns on a sphere

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(*Received* 20 *February* 1959)

Abstract—A treatment of stationary planetary flow patterns driven by source-sink distributions in a cylindrical tank (Stommel *et al.*, 1958) is extended to predict flow patterns which might be expected under similar circumstances on a rotating sphere. Flow patterns are sketched for various source-sink distributions and meridional and zonal boundary conditions.

(1) INTRODUCTION

IN a previous paper (STOMMEL *et al.*, 1958) we defined a regime of flow specified by the following elements : (a) the flow in the whole layer is steady and geostrophic except and only (b) at the western boundary where a narrow, intense boundary current is permitted to depart markedly from geostrophy, and (c) the system, which would otherwise be at rest, is driven by a distribution of sources and sinks of fluid (this might include driving agents, such as wind, which can be expressed in terms of source and sink distributions).

Such regimes can occur in a homogeneous layer of fluid (a) of uniform or varying depth on a rotating sphere (b) of uniform or varying depth on a beta plane (c) of radially non-uniform depth on a rotating plane, i.e., in a cylindrical rotating tank with a level bottom. Predictions of patterns of flow deduced from the defined regime have been verified experimentally in the cylindrical rotating tank. It is our purpose here to set forth some theoretical results with respect to circulation patterns which may be deduced for similar flows on a rotating sphere. Where these circulation patterns can be related to existing geometry and boundary conditions in the actual ocean, they might be taken as highly abstracted and idealized models of abyssal ocean circulation. Such models are, of course, highly speculative, since we do not have clear cut confirmation of the existence of the flows described, but we have been encouraged by the success of the rotating tank results to put forth a number of spherical models in order to stimulate further speculation.

In applying this model to the abyssal layer of the ocean, we visualize a distributed sink as involving a transfer of fluid *upward* through the main thermocline, and a concentrated source as a supply of sinking, cold fluid generated in a small area at high latitude.

Our justification for assuming, in these simple models, that there is a general upward flow of water — of several centimetres per day — at mid-depths, is the theory of the oceanic thermocline developed by ROBINSON and STOMMEL (1959). This theory treats each ocean basin as a separate entity. We will show in Part II of this report how a series of such basins can be, in principle, connected together to form a model of the abyssal circulation of the world ocean.

Vol
6
1959

(2) THE FORMULATION OF THE EQUATIONS FOR A HOMOGENEOUS OCEAN

Consider a globe of radius a, rotating with angular velocity ω, covered with a layer of homogeneous water of depth h. Let u be southward component of velocity, and v be eastward component of velocity at a point in co-latitude ϑ and longitude ϕ. (Fig. 1.). Assuming hydrostatic pressure in the vertical, small motions, and absence of an external disturbing force, the linearized dynamical equations are*

$$\frac{\partial u}{\partial t} - 2\omega v \cos\vartheta = -\frac{g}{a}\frac{\partial \zeta}{\partial \vartheta}$$

$$\frac{\partial v}{\partial t} + 2\omega u \cos\vartheta = \frac{-g}{a \sin\vartheta}\frac{\partial \zeta}{\partial \phi}$$

where ζ is the displacement of the free surface.

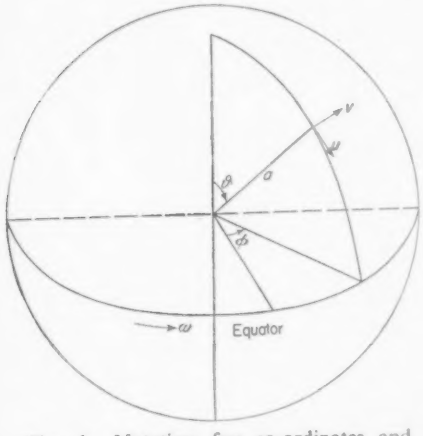

Fig. 1. Notation for co-ordinates and velocities.

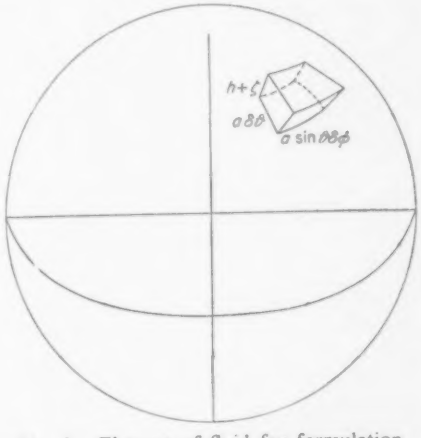

Fig. 2. Element of fluid for formulation of continuity equation.

We must now obtain a suitable form of the continuity equation. Consider a column of water shown in Fig. 2. Neglecting the quantity ζ compared to h, the rate of flow of water out of the column through the vertical walls is

$$\frac{\partial}{\partial \vartheta}(hua \sin\vartheta)\,\delta\phi\,\delta\vartheta + \frac{\partial}{\partial \phi}(hva)\,\delta\phi\,\delta\vartheta$$

At the top of the column, which is of area $a^2 \sin\vartheta\,\delta\phi\,\delta\vartheta$, the free surface is rising at the rate $\partial\zeta/\partial t$, and also there is passage of water through the free surface (positive upward) of $Q(\vartheta, \phi)$ per unit area. (We avoid GOLDSBROUGH's notation $P(\vartheta, \phi)$ because it suggests the word precipitation, of which it is indeed negative, and also looks like Legendre polynomials). We shall call Q the sink function ; it represents a distributed sink.

The total flux upward is therefore $\left(Q + \frac{\partial \zeta}{\partial t}\right) a^2 \sin\vartheta\,\delta\vartheta\,\delta\phi$. Hence the continuity equation is

*We set up the same equations as GOLDSBROUGH (1933).

$$\frac{\partial}{\partial \vartheta}(hu \sin \vartheta) + \frac{\partial}{\partial \phi}(hv) + \left(Q + \frac{\partial \zeta}{\partial t}\right) a \sin \vartheta = 0 \tag{1.2}$$

Let us now consider a steady circulation in a layer of uniform depth. In this case h is constant, and $\frac{\partial u}{\partial t} = \frac{\partial v}{\partial t} = \frac{\partial \zeta}{\partial t} = 0$. The solution of (1) and (2) is then

$$u = -\operatorname{cotan} \vartheta \cdot \frac{Qa}{h} \tag{1.3}$$

$$\frac{\partial v}{\partial \phi} = -\tan \vartheta \cdot \frac{a}{h} \cdot \frac{\partial (\cos^2 \vartheta \cdot Q)}{\partial (\cos \vartheta)} \tag{1.4}$$

$$\frac{\partial \zeta}{\partial \phi} = 2\omega \frac{a^2}{gh} \cos^2 \vartheta \cdot Q \tag{1.5}$$

An alternate form of (4) :

$$\frac{\partial v}{\partial \phi} = \frac{a}{h} \frac{1}{\cos \vartheta} \frac{\partial}{\partial \vartheta}(Q \cos^2 \vartheta)$$

$$= \frac{a}{h}\left[\cos \vartheta \frac{\partial Q}{\partial \vartheta} - 2 \sin \vartheta \cdot Q\right] \tag{1.6}$$

(3) EVAPORATION — PRECIPITATION HEMISPHERES WITHOUT MERIDIONAL BOUNDARIES

Consider the globe with no boundaries (coasts). The sink function $Q(\vartheta, \phi)$ is taken as

$$Q = Q_0 \sin \phi \sin \vartheta \tag{2.1}$$

This form of source-sink distribution corresponds to dividing the sphere into a western hemisphere ($\pi \leqslant \phi < 2\pi$) in which Q is negative (precipitation, source) and an eastern hemisphere ($0 \leqslant \phi < \pi$) in which Q is positive (evaporation, sink).

The expressions for u, v, ζ are as follows :

$$u = \frac{Q_0 a}{h} \sin \phi \cos \vartheta \tag{2.2}$$

$$\frac{\partial v}{\partial \phi} = \frac{Q_0 a}{h}[3 \cos^2 \vartheta - 2] \sin \phi \tag{2.3}$$

$$\frac{\partial \zeta}{\partial \phi} = \frac{2\omega a^2 Q_0}{gh} \cos^2 \vartheta \sin \vartheta \sin \phi \tag{2.4}$$

or when integrated

$$u = \frac{Q_0 a}{h} \sin \phi \cos \vartheta \tag{2.5}$$

$$v = \frac{Q_0 a}{h}[2 - 3 \cos^2 \vartheta] \cos \phi + \frac{g}{2\omega a \cos \vartheta} \frac{dG}{d\vartheta} \tag{2.6}$$

$$\zeta = -\frac{2\omega a^2 Q_0}{gh} \cos^2 \vartheta \sin \vartheta \cos \phi + G(\vartheta) \tag{2.7}$$

Vol 6 1959

where $G(\vartheta)$ is an arbitrary function physically associated with an arbitrary zonal flow superposable as an initial condition in the general solution.

The distribution of the sink-function and contours of disturbed height ζ, or simply isobars, are shown in Fig. 3, for $G(\vartheta) = 0$.

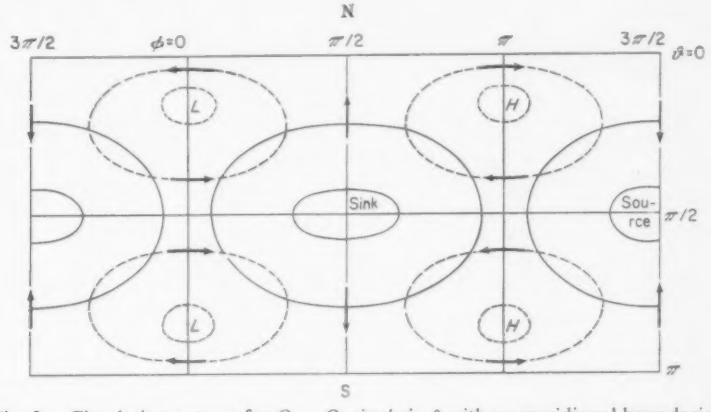

Fig. 3. Circulation pattern for $Q = Q_0 \sin\phi \sin\vartheta$ with no meridional boundaries.

Four self-contained circulation cells exist, separated by the meridian $\phi = 0, \pi$ and by the equator $\vartheta = \pi/2$. The flow is equatorward in the source hemisphere, poleward in the sink hemisphere. The geostrophic planetary flow field is divergent because the Coriolis parameter $2\omega\cos\vartheta$ is a function of latitude. Although all flow is parallel to the isobars, the transport between isobars must vary with latitude. The fields of motion which we derive by the method of GOLDSBROUGH are precisely those whose planetary-geostrophic convergence matches the distributed sink function everywhere.

If we choose the 'step' distribution corresponding to $Q = +Q_0$ on the sink hemisphere $(0 \leqslant \phi < \pi)$ and $Q = -Q_0$ on the source hemisphere $(\pi \leqslant \phi < 2\pi)$, the solutions of (1.5) and (1.6) are

$$\zeta = \frac{2\omega a^2 Q_0}{gh}\cos^2\vartheta \,.\, \phi + G_1(\vartheta), \qquad (0 \leqslant \phi < \pi)$$

$$= -\frac{2\omega a^2 Q_0}{gh}\cos^2\vartheta \,.\, \phi + G_2(\vartheta), \qquad (\pi \leqslant \phi < 2\pi) \tag{2.8}$$

$$v = -\frac{2Q_0 a}{h}\sin\vartheta \,.\, \phi + \frac{g}{2\omega a\cos\vartheta}\frac{dG_1}{d\vartheta}, \qquad (0 \leqslant \phi < \pi)$$

$$= \frac{2Q_0 a}{h}\sin\vartheta \,.\, \phi + \frac{g}{2\omega a\cos\vartheta}\frac{dG_2}{d\vartheta}, \qquad (\pi \leqslant \phi < 2\pi) \tag{2.9}$$

where G_1 and G_2 must be selected so as to provide continuity of ζ at the boundary between the two hemispheres.

A simple, symmetrical flow pattern analogous to that of Fig. 3 is obtained if we take G_1 and G_2 such as to make $\zeta = 0$ around the great circle meridian $\phi = \frac{\pi}{2}, \frac{3\pi}{2}$.

The result is :

$$\zeta = \frac{2\omega a^2 Q_0}{gh} \cos^2\vartheta \cdot \left(\phi - \frac{\pi}{2}\right), \qquad (0 \leqslant \phi < \pi)$$

$$= \frac{2\omega a^2 Q_0}{gh} \cos^2\vartheta \cdot \left(\frac{3\pi}{2} - \phi\right), \qquad (\pi \leqslant \phi < 2\pi) \tag{2.10}$$

$$v = \frac{2Q_0 a}{h} \sin\vartheta \cdot \left(\frac{\pi}{2} - \phi\right), \qquad (0 \leqslant \phi < \pi)$$

$$= \frac{2Q_0 a}{h} \sin\vartheta \cdot \left(\phi - \frac{3\pi}{2}\right), \qquad (\pi \leqslant \phi < 2\pi) \tag{2.11}$$

and u is given by (1.3) with appropriate substitution of $\pm Q_0$ for Q. The circulation pattern is sketched in Fig. 4. There is a singularity at the poles where the velocity u is infinite. Since these conditions are contradictory with the original requirement that the motions be small, we can expect that a higher order dynamical system will be required to explain the detailed configuration at the poles themselves. This points up some of the difficulties which actually could occur in applying the GOLDSBROUGH method to a simple experimental or geophysical system. These difficulties can be avoided, of course, if we are able to use certain simple forms of sink function such

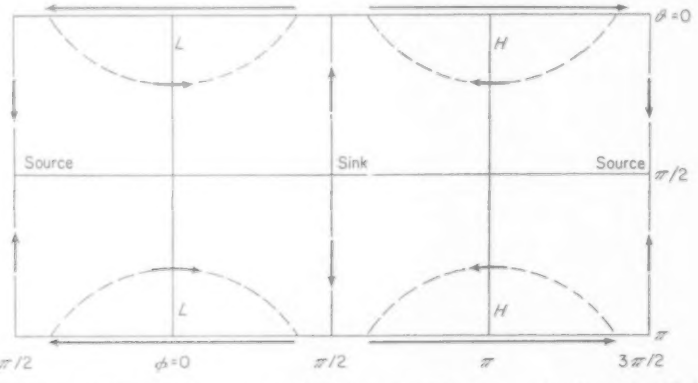

Fig. 4. Circulation pattern for $Q = +Q_0$, $(0 \leqslant \phi < \pi)$, and $Q = -Q_0$, $(\pi \leqslant \phi < 2\pi)$, with no meridional boundaries.

as that employed in Fig. 3. To indicate in Fig. 4 the need for an additional higher order regime at the poles, we resort to a simple symbolic device: We close the isobars in a narrow boundary region at the poles as shown in Fig. 4, so that we do not represent a mathematical discontinuity in pressure, and we draw a heavy arrow at the top to indicate the direction of the singular flow over the poles in this region. The arrows over the lows and highs are both in the same direction, the projection makes them look different. On the sphere, looking at the North Pole, the picture looks something like that shown in Fig. 5.

For simplicity of expression, let us call this little intense stream, the ' polar jet,' or ' polar pinch.'

The polar jet, perhaps surprisingly at first glance, flows from the sink hemisphere to the source hemisphere.

Vol 6 1959

The transport of water in the polar jet must be

$$\frac{gh}{2\omega}(\zeta_{\max} - \zeta_{\min}) \tag{2.12}$$

From equation (2.8) we find

$$\zeta_{\max} - \zeta_{\min} = \frac{2\omega a^2}{gh}\pi Q_0 \tag{2.13}$$

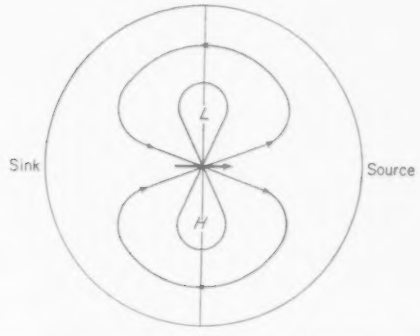

Fig. 5. Circulation pattern of Fig. 4 as seen looking down on the North Pole.

Thus the transport of both the polar jets into the source hemisphere is

$$2\pi a^2 Q_0 \tag{2.14}$$

The total integrated source flowing into the source hemisphere is

$$2\pi a^2 Q_0.$$

Therefore, in the steady state there must be a relatively massive flow, $4\pi a^2 Q_0$, flowing geostrophically from the source hemisphere into the sink hemisphere across the great circle — meridian — bounding the two hemispheres. We may verify this by the following direct calculation: From (2.11) we have for the total zonal flux, F, of material out of the source hemisphere along the great circle meridian $\phi = 0, \pi$:

$$F = 2\pi a\, Q_0 \sin\vartheta.$$

The total zonal flow out of the source hemisphere is then

$$\int_0^{\pi} 2\pi a^2 Q_0 \sin\vartheta\, d\vartheta = 4\pi a^2 Q_0,$$

in agreement with the value calculated above from the sum of the transport in the polar jet and the distributed source.

(It might be noted that a somewhat different flow pattern with similar polar jet and zonal transports is obtained if one selects G_1 and G_2 in (2.8) in such a way as to make $\zeta = 0$ along the great circle meridian $\phi = 0, \pi$).

(3) BASINS BOUNDED BY MERIDIANS

We now proceed to discuss the case where an ocean basin is bounded by meridians ϕ_1 and ϕ_2 extending all the way to the poles. Here we develop the pressure pattern from the eastern boundary ϕ_2 where, of course, there can be no zonal transport,

so we take $\phi = \phi_2$ as an isobar : $\zeta = 0$. As before, we require geostrophic flow in the interior, but we shall now satisfy continuity requirements by allowing departure from geostrophy in a western boundary current. Our justification for the treatment carried out below rests on the success with which the same method predicts and describes similar flow patterns in a rotating cylindrical tank (STOMMEL *et al.*, 1958).

As an initial illustration we take a uniformly distributed sink Q_0 over the entire surface and balance this by a concentrated source S_0 at the pole, considering the basin north of the equator for simplicity (Fig. 6). Since the surface area of the sector is $a^2(\phi_2 - \phi_1)$, we take the concentrated source at the north pole as $S_0 = Q_0 a^2 (\phi_2 - \phi_1)$. The geostrophic velocities and surface isobars are given by :

$$u = -\frac{Q_0 a}{h} \operatorname{cotan} \vartheta$$

$$v = \frac{2Q_0 a}{h} \sin \vartheta \,.\, (\phi_2 - \phi)$$

$$\zeta = \frac{2\omega a^2 Q_0}{gh} \cos^2 \vartheta \,.\, (\phi - \phi_2) \tag{3.1}$$

The pattern of the isobars is shown in Fig. 6. (There is a singularity at the pole similar to that encountered in Section 2).

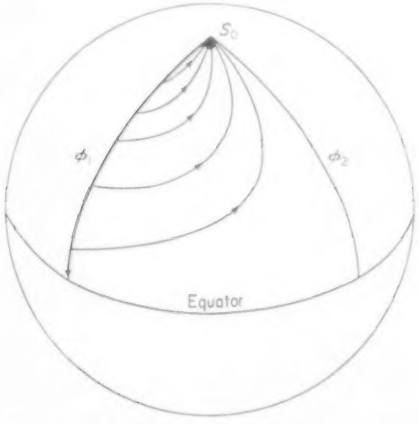

Fig. 6. Circulation pattern in meridionally bounded ocean with concentrated source S_0 at North Pole and a uniformly distributed sink Q_0, such that $S_0 = Q_0 a^2 (\phi_2 - \phi_1)$.

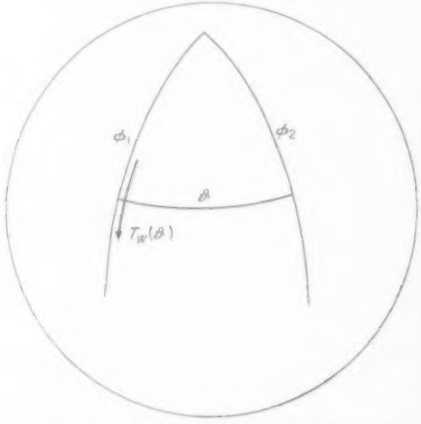

Fig. 7. Notation for evaluating strength of western boundary current $T_W(\vartheta)$ by consideration of continuity of mass flow in a sector.

This flow does not satisfy boundary conditions at $\phi = \phi_1$, and, following the procedure used by (STOMMEL, *et al.*, 1958) we introduce symbolicaly a western boundary current $T_W(\vartheta)$ shown positive southward by the heavy arrow in Fig. 7.

Consider a section of the ocean (Fig. 7) bounded by ϕ_1, ϕ_2, and ϑ. The total outward flux of mass from portions of the basin so bounded consists of four parts :

1. The western boundary current $T_W(\vartheta)$.

2. The flux integrated over the surface area of the basin

$$= \int_0^{\vartheta}\int_{\phi_1}^{\phi_2} Q\, a^2 \sin\vartheta \,.\, d\vartheta\, d\phi.$$

3. The flux across ϑ in the interior of the ocean, away from the western boundary current

$$\int_{\phi_1}^{\phi_2} ahu \sin\vartheta \, d\phi.$$

4. The flux $-S_0$ from any concentrated source present in the section.

In the stationary state these must add up to zero, hence

$$T_W(\vartheta) = -\int_{\phi_1}^{\phi_2} ahu \sin\vartheta \, d\phi - \int_0^{\vartheta}\int_{\phi_1}^{\phi_2} Qa^2 \sin\vartheta \, d\vartheta \, d\phi + S_0 \tag{3.2}$$

Evaluating equation 3.2 in the light of 3.1, we obtain for case of Fig. 6 :

$$T_W = 2Q_0 a^2 (\phi_2 - \phi_1) \cos\vartheta \tag{3.3}$$

$$= 2S_0 \cos\vartheta.$$

Thus in the neighbourhood of the pole the transport in the western boundary current, $T_W(0) = 2Q_0 a^2 (\phi_2 - \phi_1) = 2S_0$, is *twice* the transport associated with the concentrated source alone, and, as in the cylindrical case developed by STOMMEL *et al.* (1958) a large volume of water is continually being *recirculated* in such a steady-state system.

Equation (3.3) however, reveals a very interesting difference between the spherical and the cylindrical geometries. Where, in the cylindrical basin, the western boundary current does not vary with radial position and, flows undiminished all the way to the outer rim of the tank, the current in the spherical geometry diminishes as it flows to lower latitudes (Fig. 6), dropping to zero at the equator. (Of course, if we assumed a concentrated source S_0 at the pole *larger* than the internal distributed sink flux $Q_0 a^2 (\phi_2 - \phi_1)$, the excess would flow across the equator to feed the sector on the other side).

The internal consistency of equations (3.1) and (3.3) is readily checked by evaluating the total zonal transport at $\phi = \phi_1$ and showing it to be equal to the initial total western boundary current $T_W(0)$, i.e.

$$\int_0^{\pi/2} hau(\phi_1)\, d\vartheta = 2Q_0 a^2 (\phi_2 - \phi_1)$$

A somewhat different pattern of western boundary currents arises if we alter the situation in Fig. 6 only by transferring the concentrated source from the pole to the southwest corner of the sector at the equator (i.e. this could stem from a western boundary current crossing the equator from the southern side). We still assume that the source strength S_0 just balances the distributed sink $Q_0 a^2 (\phi_2 - \phi_1)$.

If we now evaluate $T_W(\vartheta)$ from equation (3.2), remembering that there is no longer a concentrated source within the section being considered in Fig. 7, we obtain :

$$T_W(\vartheta) = Q_0 a^2 (\phi_2 - \phi_1)(1 - 2\cos\vartheta) \tag{3.4}$$

The interior flow pattern is still defined by (3.1) exactly as it was in Fig. 6, but the nature of the western boundary current is very different and is sketched in Fig. 8.

A current of recirculated water equal to the source strength starts at the pole and flows toward the source just as in the cylindrical case (STOMMEL *et al.*, 1958), but in the spherical geometry this current gradually diminishes to zero at $\vartheta = 60°$ (30° north latitude). A northward current of equal strength starts at the equatorial source

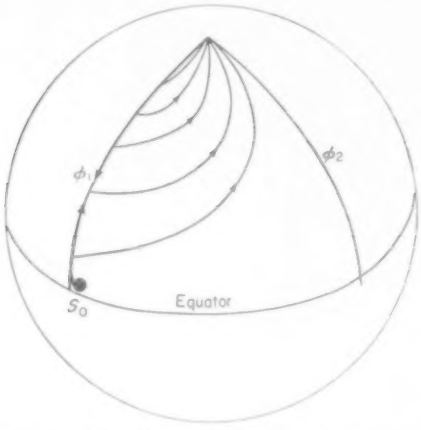

Fig. 8. Circulation pattern in meridionally bounded ocean with concentrated source S_0 (fed by western boundary current from below the equator) and a uniformly distributed sink Q_0 such that $S_0 = Q_0 a^2 (\phi_2 - \phi_1)$.

and also diminishes to zero at 30° north latitude. (This feature is completely absent in the corresponding cylindrical situation). The zonal transport out of these two currents feeds exactly the same interior flow pattern as in Fig. 6.

In the second paper of this series, we shall repreatedly make use of the above results in order to sketch idealized patterns of boundary currents and interior circulation in various basins of the world ocean.

(4) FURTHER ILLUSTRATIONS OF CIRCULATIONS IN MERIDIONALLY BOUNDED BASINS

Applying the analytical method outlined in Section 3, we now state the results obtained for a few other basically interesting cases of source-sink distributions.

(a) A simple form of the function Q which avoids singular flows at the poles is

$$Q = Q_0 \sin^n \vartheta \cos \vartheta, \qquad n \geqslant 1 \tag{4.1}$$

Thus the half of the basin north of the equator is a sink while the southern half is an equal source. S_0 is taken as zero.

The isobars are given by the disturbed height

$$\zeta = \frac{2\omega a^2 Q_0}{gh} \sin^n \vartheta \cos^3 \vartheta \,.\, (\phi - \phi_2) \tag{4.2}$$

and the western boundary current transport by

$$T_W(\vartheta) = Q_0 a^2 \frac{\sin^n \vartheta}{n+2} [(n+3)\cos^2 \vartheta - 1] \,.\, (\phi_2 - \phi_1). \tag{4.3}$$

Vol 6 1959

The interior isobars and nature of the western boundary current are sketched for the case $n = 0$ in Fig. 9. The north and southbound boundary currents fall to zero at $\vartheta = 54{\cdot}7°$. Here the flow has a singularity at the poles. The singularity is eliminated when $n \geqslant 1$. For larger values of n the flow is essentially similar to that sketched in Fig. 9 with isobars somewhat more concentrated in mid-latitudes.

(b) It is interesting to examine the effect of a sloping bottom, taking $h = h_0 \sin \vartheta$. Here we must go back to the continuity equation in the form of (1.2). For a uniformly distributed sink $Q = Q_0$ and a concentrated source in the south, we obtain :

$$u = -\frac{Q_0 a}{h_0} \cos \vartheta$$

$$v = \frac{Q_0 a (\phi - \phi_2)}{h_0} (1 - 3 \sin^2 \vartheta)$$

$$\zeta = \frac{Q_0 a^2 \omega}{g h_0} (\phi - \phi_2) \cos \vartheta \sin^2 \vartheta$$

$$T_W(\vartheta) = Q_0 a^2 (\phi_2 - \phi_1) [\cos \vartheta (1 + \sin^2 \vartheta) - 1] \tag{4.4}$$

The isobars and western boundary transport are sketched in Fig. 10. Comparing with Fig. 6, we see that the effect of the sloping bottom is to cause the isobars to intersect the ϕ_1 meridian at different latitudes instead of requiring a common intersection at the pole as in the case of uniform depth. It should also be noted that the southward flowing boundary current first increases in intensity and then decreases instead of showing a monotonic decrease.

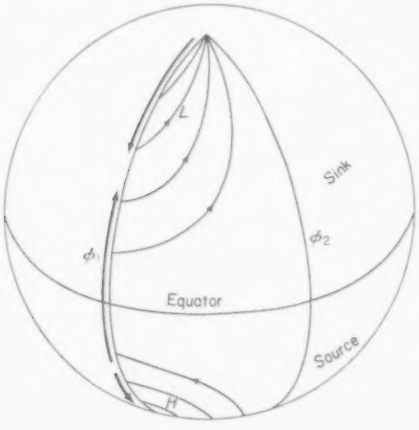

Fig. 9. Circulation pattern for $Q = Q_0 \cos \vartheta$ with no concentrated sources. Note boundary current crossing equator.

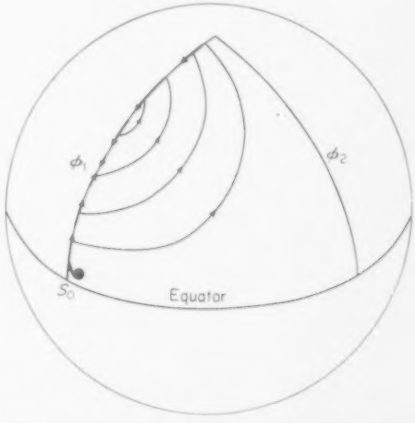

Fig. 10. Circulation pattern in meridionally bounded ocean with sloping bottom. $h = h_0 \sin \vartheta$. Concentrated southern source and uniformly distributed sink.

(c) In any source-sink distribution in which Q becomes zero along a latitude circle, the interior geostrophic flow does not cross this latitude circle, and the circulation in a basin is divided up into separate cells. As an illustration we take

$$Q = -Q_0 \cos 2\vartheta \tag{4.5}$$

giving a distributed source in latitudes higher than 45° and a distributed sink in the belt between the 45° latitude circles. Since the source flux is smaller than the sink flux, we balance the flux by introducing concentrated sources $S_0 = \frac{1}{3} Q_0 a^2 (\phi_2 - \phi_1)$ at each pole. In this case we obtain for a basin of uniform depth :

$$u = \frac{Q_0 a}{h} \cdot \frac{2\cos^3\vartheta - \cos\vartheta}{\sin\vartheta}$$

$$v = \frac{2Q_0 a}{h} (\phi - \phi_2) \sin 3\vartheta$$

$$\zeta = -\frac{2Q_0 a^2 \omega}{gh} (\phi - \phi_2) \cos 2\vartheta \cos^2\vartheta$$

$$T_W(\vartheta) = \frac{Q_0 a^2}{3} (\phi_2 - \phi_1) \cos\vartheta (3 - 4\cos^2\vartheta) \tag{4.6}$$

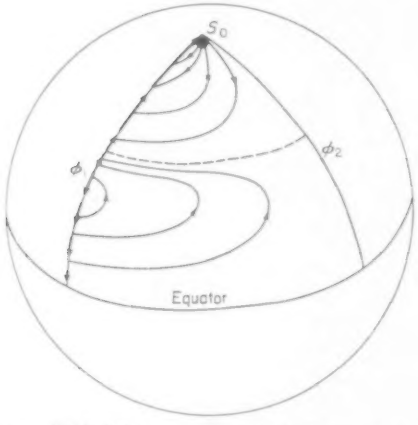

Fig. 11. Circulation pattern divided into separate cells when $Q = 0$ along a latitude circle. $S_0 = \frac{1}{3} Q_0 a^2 (\phi_2 - \phi_1)$; $Q = -Q_0 \cos 2\vartheta$.

The isobars and western boundary transports are sketched in Fig. 11, showing the cells into which the circulation is divided.

(5) BASINS BOUNDED BY MERIDIANS ϕ_1, ϕ_2 AND A NORTHERN LATITUDE ϑ_1 (FIG. 12)

For any distribution of Q, the expressions for u, v and ζ are the same as before, except now it will, in general, be necessary to add a boundary layer at both $\vartheta = \vartheta_1$ and $\phi = \phi_1$. The expression for $T_W(\vartheta)$ must be changed because there is no loss from the sink defined north of ϑ_1 ;
thus

$$T_W(\vartheta) = -\int_{\phi_1}^{\phi_2} ahu \sin\vartheta \, d\phi - \int_{\phi_1}^{\phi_2}\int_{\vartheta_1}^{\vartheta} Q a^2 \sin\vartheta \, d\vartheta \, d\phi \tag{5.1}$$

As an example, we may choose $Q = Q_0$, a constant. Hence

$$T_W(\vartheta) = Q_0 a^2 (\phi_2 - \phi_1)(2\cos\vartheta - \cos\vartheta_1) \tag{5.2}$$

Vol 6 1959

The critical latitudes, where the western current reverses direction, are at

$$\vartheta = \cos^{-1}\left(\tfrac{1}{2}\cos\vartheta_1\right) \tag{5.3}$$

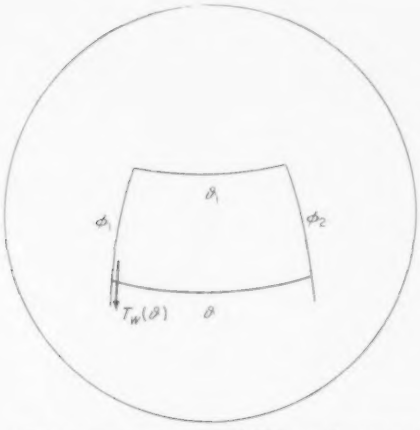

Fig. 12. Notation for sector with rigid boundary at co-latitude ϑ_1.

In the limiting case where the latitude of the northern boundary approaches the pole ($\vartheta_1 \to 0$) the latitude of reversal of the western boundary current is about ($\vartheta = 60°$) 30° N lat.

In some actual applications an additional point source S_0 should be added in the basins, near the poles, to supply the material lost by the distributed sink. In this case we must add S_0 to the equation on the right hand member.

(6) COMPLETELY CLOSED BASIN BOUNDED BY MERIDIANS ϕ_1, ϕ_2 AND LATITUDE CIRCLES ϑ_1, ϑ_2 (FIG. 13)

Again, we take a distributed sink function $Q = Q_0$. The quantities u, ζ, $T_W(\vartheta)$ are the same as in the case of Section 5, but, of course, for a steady state there must be a source of water somewhere. We place it at the point ϕ_1, ϑ_2. Its intensity must be equal to that of the distributed sink over the whole ocean area.

$$S_0 = \int_{\phi_1}^{\phi_2}\int_{\vartheta_1}^{\vartheta_2} Qa^2 \sin\vartheta \, d\vartheta \, d\phi = Q_0 a^2 (\phi_2 - \phi_1)(\cos\vartheta_1 - \cos\vartheta_2) \tag{6.1}$$

The isobar pattern is given by

$$\zeta = \frac{2Q_0 a^2 \omega}{gh}(\phi_2 - \phi)\cos^2\vartheta \tag{6.2}$$

the pressure pattern being developed from the eastern boundary with zero zonal transport at $\phi = \phi_2$ as before.

In Fig. 13 the isobar pattern according to (6.2) is drawn. It is seen that boundary currents must be drawn at all except the eastern boundary. The western boundary current flows northward from the source, and across the equator, and is met in middle northern latitudes by a southward flowing western boundary current. For example, if we take the latitude of the northern boundary of the ocean 70°N, the latitude of

the reversal of the western boundary current is 18·5°N ($\vartheta = 81{\cdot}5°$) from equation (5.3).

If, in the configuration shown in Fig. 13 we now put another point source S_1, at the north-western corner of the basin, and assert that there is a distributed sink whose intensity integrated over the entire area of the ocean is equal to the sum of the two point sources : $S_0 + S_1$, we get a variety of circulation patterns, depending on the relative strengths of sources S_0 and S_1. If they are equal, the flow is in two

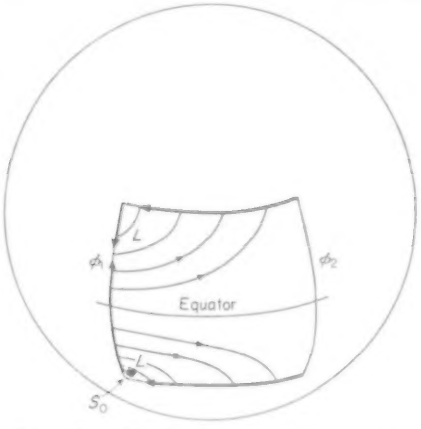

Fig. 13. Circulation pattern in sector bounded by meridians and latitude circles. Boundary currents will be present along north and south boundaries as well as along the western boundary. Single concentrated source.

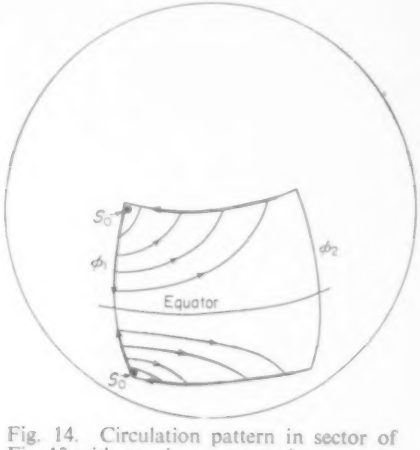

Fig. 14. Circulation pattern in sector of Fig. 13 with equal concentrated sources in each hemisphere.

separate gyres with no flow across the equator (Fig. 14). If the sources are not equal there is a flow across the equator in the western boundary current (there is never transequatorial flow in the interior) from the basin with the larger source, the pattern then looking more like that shown in Fig. 13. The only difference in these patterns, as we vary the relative strength of S_0 and S_1 is in the latitude of reversal of the western boundary current.

(7) SOME COMMENTS ON THE VARIOUS BOUNDARY LAYERS

The various boundary layers introduced in the preceding examples can be avoided, formally, by proper choice of the sink function Q. Thus, all boundary currents on latitude circles can be avoided by choosing Q so that it approaches zero at the boundary. Similarly, western boundary currents can be avoided if we restrict $Q(\vartheta, \phi)$ to forms which satisfy the integral relation

$$\int_{\phi_1}^{\phi_2} Q(\vartheta, \phi)\, d\phi = 0$$

where ϕ_2 and ϕ_1 are the longitudes of the meridional boundaries of the basin. In effect, these are the restrictions which GOLDSBROUGH imposes in his investigation. Such restrictions are, however, quite artificial from a physical point of view. Most

simple physical regimes in experiments, such as those in the rotating circular basin with parabolic depth law (STOMMEL *et al.*, 1958) as well as geophysical or oceanographic situations, are certainly not bound to conform to such artificial restrictions. Thus, there is a need to take a rather broad view in choosing models with various forms of Q, and this evidently requires introduction of appropriate boundary layers.

(8) PARTIAL BARRIERS

What will the regime of flow look like when there are meridional barriers which end abruptly at some latitude far from any other coast ? Let us return to the simple configuration originally drawn in Fig. 7, but now we put a long barrier along the longitude ϕ' ($\phi_1 < \phi' < \phi_2$) extending from the pole to colatitude ϑ' but not beyond it (Fig. 5).

We can make use of the same kind of reasoning we used before to construct interior solutions in the open ocean portion. However it is clear that we can now have two western boundary currents along ϕ_1 and ϕ', with transports T_W and T_W'. If upon calculation of the transport T_W' along the partial barrier ϕ' it is found that $T_W' \neq 0$ at $\vartheta = \vartheta'$, then clearly the only resort is a zonal current (Z in Fig. 16) to connect

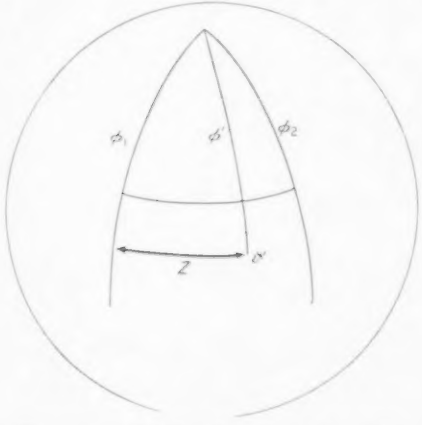

Fig. 15. Meridionally bounded sector (ϕ_1, ϕ_2) with partial meridional barrier extending from pole to longitude ϕ'.

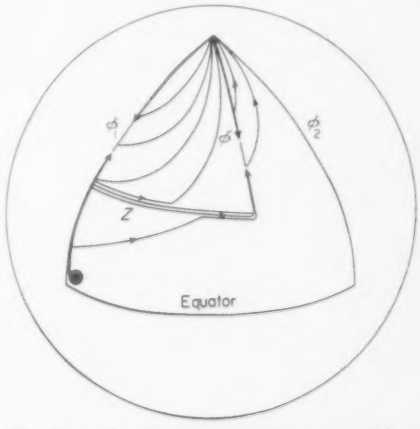

Fig. 16. Circulation pattern in sector with partial meridional barrier. Western boundary currents along ϕ_1 and ϕ' are connected by zonal flow Z.

the western boundary current on the barrier ϕ' to that on the barrier ϕ_1. In this way continuity can be preserved. To see how this alters the picture of the circulation we can compare Figs. 8 and 16. The former shows the pattern of circulation for a basin, bounded by two meridians and the equator, filled by a source S_0 at the equator, and emptied by a uniformly distributed sink over the whole oceanic area. In Fig. 16 a meridional barrier ϕ' has been added part way down from the pole, and the mass flux into the eastern basin occurs partly as a zonal current toward the east. The fact that this zonal current can indeed ' round the corner ' to join the western boundary current has been demonstrated in rotating tank experiments by A. J. FALLER (private communication). Both north and south of the zonal current there is an interior meridional component of flow which must join properly to the zonal current.

Therefore the zonal current is not strictly speaking a continuous current ; water is added from the south and thrown off to the north along its length, as the drawing of the isobars indicates.

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Contribution No. 1023 *Woods Hole Oceanographic Institution. Part of this work was done while one of us (A.B.A.) was the recipient of a Fellowship of the John Simon Guggenheim Memorial Foundation.*

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Vol
6
1959

Some measurements of deep currents in the Eastern North Atlantic

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Abstract—Deep current measurements extending over several weeks were made in a small area of the Eastern North Atlantic during the summer of 1958, and a number of hydrographic stations were occupied in the same area. The work formed part of the programme of a joint British-Norwegian I.G.Y. cruise. The currents were found to be variable in time with periods of a few weeks, and in space over a distance of a few tens of miles. In a vertical section, velocities did not decrease uniformly with depth, and no ' level of no motion ' was found.

INTRODUCTION

FEW direct measurements of currents have been made in deep water, and none has been continued for longer than a week. Comparisons between the results of such measurements, and the water movements deduced from geostrophic calculations or from a study of water-mass characteristics, have often been unsatisfactory (BOWDEN, 1954). The aim of the work described here was to extend the period of observation of currents in one area to several weeks, to see whether any long-period fluctuations could be detected, and possibly to arrive at a mean current more suitable for comparison with the indirect observations. An attempt was also made to study the spatial variability of currents at a given depth, by making simultaneous observations several miles apart.

The choice of area was controlled by several factors. A fairly flat sea-floor seemed desirable, so that the currents might be assumed to be unaffected by irregularities of topography. It seemed preferable to work where weak currents might be expected, so that the distance moved by the water in a few weeks might not be excessive ; and, most important, good weather was needed.

The neighbourhood of 41½°N, 14½°W seemed most suitable (Fig. 1). It is near the western edge of the Iberian abyssal plain. There are several small outcrops rising a few hundred metres above the level of the plain, which were useful as fixed reference marks for navigation but which seemed too small to have much influence on the water movements. The area was also of some interest for geophysical investigations on the sea floor, which were made during the cruise by members of the Geophysics Department of the University of Cambridge.

Most of the water-sampling in this area was done by the M.S. *Helland-Hansen*, and the current-measurements described here were made by the R.R.S. *Discovery II*. A rectangular pattern of four hydrographic stations was also worked by the R.R.S. *Discovery II* in the earlier part of the cruise.

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CURRENT-MEASURING EQUIPMENT AND METHOD

Neutrally-buoyant floats were used for the current measurements (SWALLOW, 1955, 1957). The acoustic transmitter was basically unchanged, but modifications were made to increase the life and the signal strength. A larger size of aluminium alloy tube was used (3½in. (9 cm) O.D. × 0·276in. (7 mm) nominal wall thickness)

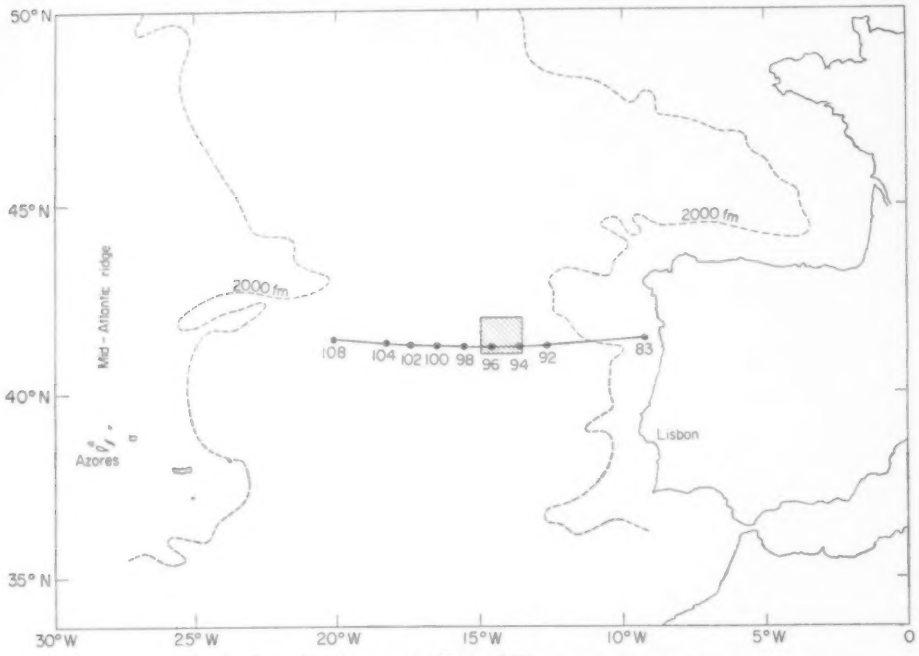

Fig. 1. Locality chart, with *Helland-Hansen* east-west section.

to accommodate more bulky components. Some improvement in signal was obtained by arranging for the units to float with the transducer at the top, and fitting them with an annular reflector plate of aluminium, 9 in. (23 cm) O.D. × ½ in. (13 mm) thick. Four different types of float were used (see Table 1).

(a) *Long-life floats* (*B*, *D*)

These carried enough batteries to run the transmitter continuously for two weeks, but the life was increased to 12 weeks by switching the transmitter on for only four hours per day, by means of a battery-driven clock fitted with a cam and microswitch. The life of the clock batteries exceeded three months. Two 8 ft (244 cm) lengths of HE-10-WP alloy tube were needed to provide enough buoyancy.

(b) *Medium-life floats* (2, 3, 5)

For these and the next two types of float, a single 12 ft (366 cm) length of RR-77 alloy tube was used. Batteries sufficient for one week's continuous running were provided, but this was increased by switching on the transmitter for four hours in every twelve by means of a spring-driven clock, which would run for about 19 days.

Vol 6 1959

(c) *Repeaters* (7)

An attempt was made to provide for more accurate depth determinations on some of the floats, by fitting them with repeaters. These consisted of a 4-stage transistor amplifier working from the transducer (used as a hydrophone) and connected so as to trigger the transmitter when a signal was received from the ship's echo-sounder. The narrowness of the beam of the echo-sounder made it necessary to locate the ship over the float with great accuracy, and only one measurement was attempted with this type of float. Besides responding to signals from the echo-sounder, the transmitter in the float sent out signals at intervals of a few seconds, so that it could be tracked in the usual way.

(d) *Short-life floats* (1, 6, 11)

These units were similar to the medium-life floats, with the spring-driven clock omitted. They would run continuously for about a week.

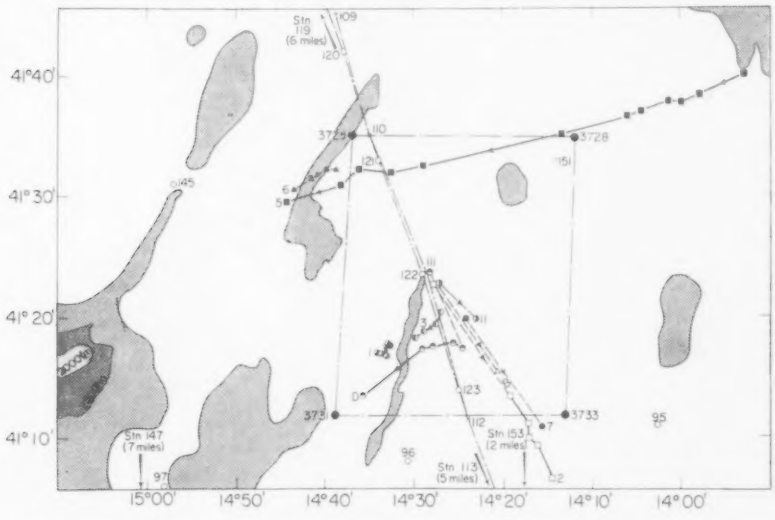

Fig. 2. Tracks of the floats, and positions of hydrographic stations. The dashed sections of the float tracks are shown on a larger scale in Fig. 3.

The method of operation was the same as that described previously (SWALLOW, 1955), fixes being made by taking groups of three bearings on a float, from positions determined by radar from anchored buoys. No aids to navigation, such as Loran or Decca, were available in the chosen area, and any movement of the anchored buoys was detected by reference to known outcrops or 'banks' rising from the abyssal plain. Buoys were located relative to each other by dead reckoning, the log being checked against radar ranges at the beginning and end of each run. Many astronomical fixes were made on buoys anchored near the northern end of the bank at 41° 23′N, 14° 29′W (Fig. 2) and it is believed that the uncertainty in absolute position of this feature is less than 2 km. The uncertainty of relative positions is believed to be less than 1 km, and in many cases less than $\frac{1}{2}$ km.

OBSERVATIONS

Nine floats were used in making the current measurements. Their tracks are shown in Figs. 2 and 3, together with the positions of the hydrographic stations and the main topographic features of the area. The abyssal plain has a mean depth of 2920 fm (5340 m) and a mean slope downwards of 1 : 4000 towards SSE. Neither of the two banks crossed by the tracks of the floats exceeds 280 fm (512 m) in height above the surrounding plain. The nearest seamount is at 41° 14′N, 15° 13′W, with a minimum depth of 1450 fm (2650 m). The eastern slopes of this feature are indicated by the 2000 and 2500 fm contours in Fig. 2.

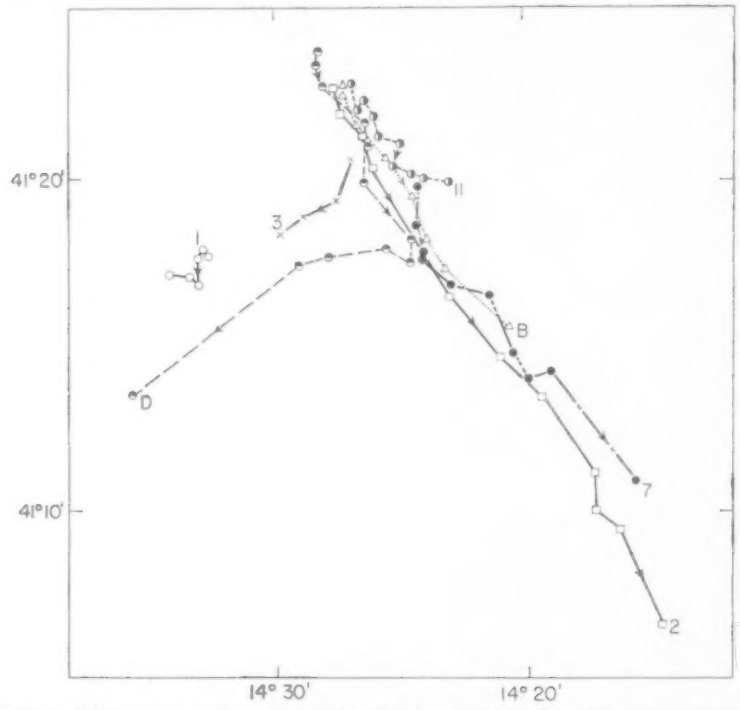

Fig. 3. Tracks of floats 1, 2, 3, 7, 11, B and D. The first position plotted for float 3 is the mean of the first four fixes.

Most of the observations were made in the two periods, 16th May-1st June, and 9th-28th June, with another visit during 9th-13th July. The main purpose of the work in the first period was to determine the variation of current with depth, and in particular to find whether a level of no horizontal motion existed. Five floats, loaded for depths between 1500 m and 4500 m, were released within 4 km radius of 41° 22′N, 14° 27′W. For practical reasons the release dates were spread out from 16th to 26th May.

The eastward and southward components of displacement of the floats are plotted against time in Fig. 4. Distances have been measured from the northern end of the adjacent bank, which was used as the fixed reference point in tracking these floats. Straight lines have been fitted to the displacements, and mean velocities computed by combining the slopes of these lines. The magnitudes and directions of the resultant

Table 1. *Summary of float observations*

Float No.		B	2	11	7	D	5	3	6	1
Nominal depth (m)		3120	2000	4500	1500	2600	2680	2600	2210	2600
Observed depth (m)		3680 ± 100	2120 ± 50	4240 ± 160	1560 ± 50[m]	2940 ± 70	2590 ± 70	2460 ± 50	2430 ± 130	2760 ± 110
Time and date laid		0011/16/V	0425/20/V	0933/23/V	1807/23/V	2315/25/V	2130/10/VI	0625/10/VI	1241/25/VI	2145/9/VII
Time and date of last fix		0230/23/V	0549/29/V	2348/28/V	0431/29/V	1930/12/VII	1115/28/VI	0800/25/VI	1115/28/VI	1232/13/VII
Velocity	Mean current (cm/sec)	2·62 ± 0·07[a]	4·74 ± 0·06[b]	1·90 ± 0·13[b]	4·14 ± 0·11[a]	1·2[c] 1·0[d]	5·19 ± 0·13[e] 4·62 ± 0·15[f]	0·1[g] 0·9[h]	3·0[j]	1·2[k] 1·3[l]
	Bearing (°true)	145·4 ± 1·4[a]	145·9 ± 0·8[b]	141·5 ± 3·7[b]	137·4 ± 1·6[a]	156[c] 233[d]	252·6 ± 1·4[e] 255·0 ± 3·3[f]	180[g] 236[h]	208[j]	187[k] 290[l]

(a) fitted by least squares
(b) fitted by least squares, but omitting last fix
(c) 0040/26/V–0103/1/VI, fitted by eye
(d) 2100/26/VI–1930/12/VII, fitted by eye
(e) 2340/10/VI–1110/17/VI, fitted by least squares
(f) 2305/22/VI–1115/28/VI, fitted by least squares.
(g) 0910/10/VI–1920/13/VI, fitted by eye
(h) 0700/17/VI–0800/25/VI, fitted by eye
(j) fitted by eye
(k) 0040/10/VII–2300/11/VII, fitted by eye
(l) 2300/11/VII–1232/13/VII, fitted by eye
(m) observed depth, by E/S, was 1420 m.

velocities are given in Table 1. Standard deviations are also tabulated, where the least-squares method was used, but in cases where few fixes were taken or an obvious curvature existed, velocities were determined from straight lines fitted by eye, and the standard deviation has not been estimated.

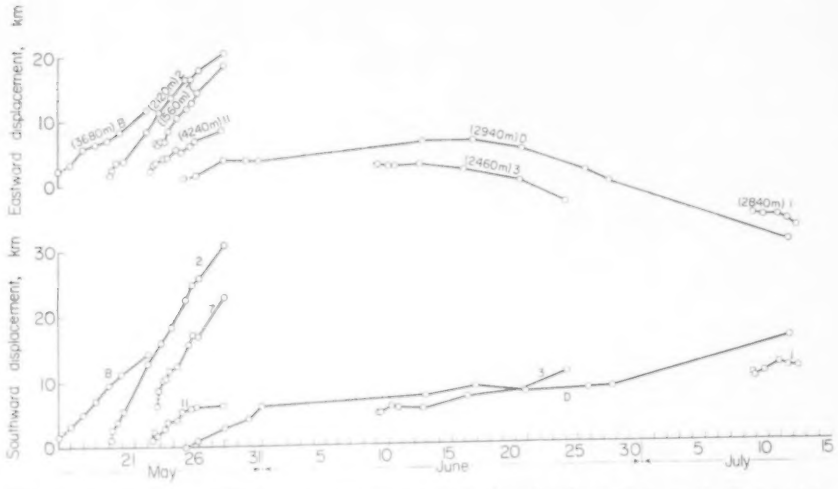

Fig. 4. Components of displacement plotted against time for floats B, 2, 7, 11, D, 3 and 1.

It is apparent, from Figs. 2, 3 and 4 and Table 1, that in the period 16th May-1st June all five floats moved in nearly the same direction (about 143°T) but with different speeds. There was no evidence for a layer of no horizontal motion, nor was there a uniform decrease of velocity with depth. Two of the five floats released, B and D, were of the long-life design, but float B developed a fault in its transmitter, the repetition rate decreasing, until after 23rd May no signals could be heard. Float D was left still working on 1st June.

An unsuccessful search was made for it on 9th-10th June, on returning after a visit to Lisbon and some geophysical work in other areas. Two medium-life floats were then launched, No. 3 near the last known position of float D, and No. 5 near a bank about 60 km to the north-east. Both were loaded for the same depth as float D, 2600 m. For the first few days, float No. 3 scarcely moved at all, and it was clear that the previous movement towards SSE had stopped. A new estimate was then made for the probable position of float D, which was found again on 13th June. Floats D, 3 and 5 were then followed for another two weeks, a further float (No. 6) being laid near No. 5 on 25th June. This new float was loaded for 2000 m and, although followed for only a short time, confirmed the direction of movement of No. 5, although its speed was rather less. The components of displacement of floats 5 and 6 are plotted against time in Fig. 5, and the tracks are shown in Fig. 2. There is a striking difference between the movements of floats 3 and 5 ; their speeds differ by a factor of five or more even though they were at one time less than 25 km apart. The average direction of movement had also shown a remarkable change, and was nearly at right angles to the previous flow.

Vol 6 1959

On returning to the area for the third visit, a short-life float, No. 1, was released near the expected position of float D. The latter was found again after three days' searching and tracking float No. 1.

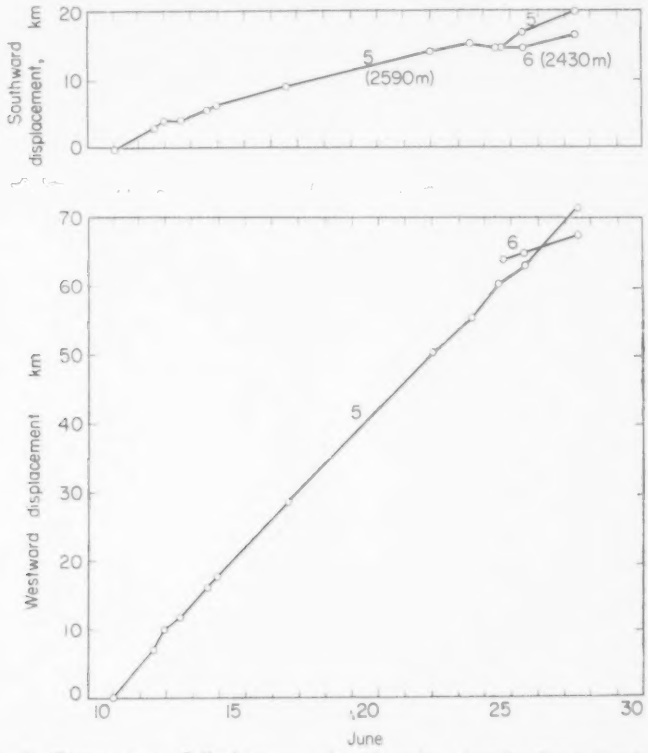

Fig. 5. Components of displacement plotted against time for floats 5 and 6.

Float D was followed for a total period of 48 days. In the first six days its velocity was approximately 1·2 cm/sec towards 156°. It must have slowed down almost immediately afterwards, and remained nearly stationary for the next 22 days. Thereafter its velocity increased again to 1·0 cm/sec, but in a direction almost at right angles to the previous movement.

The observed mean depths of the floats, together with the intended depths are given in Table 1. The two long-life floats, B and D, both went much deeper than had been intended. The other floats show a mean discrepancy of ± 150 m from the intended depth, which is not much greater than the mean of the standard deviations of the observed depths (± 100 m). The cause of the unexpected deepness of floats B and D is not known. It is probably not due to leakage since the depth determinations on float D show no significant increase with time.

COMPARISON WITH GEOSTROPHIC CURRENTS

Four hydrographic stations (3725, 3728, 3731 and 3733, (Fig. 2)) were occupied by the R.R.S. *Discovery II* between 15th and 19th May, when only float B was being followed. Temperatures and salinities were observed at 24 depths between the

surface and 5000 m at all stations except 3725, where no observations were obtained below 3500 m. Salinities were measured on board by an electrical conductivity method (Cox, 1958).

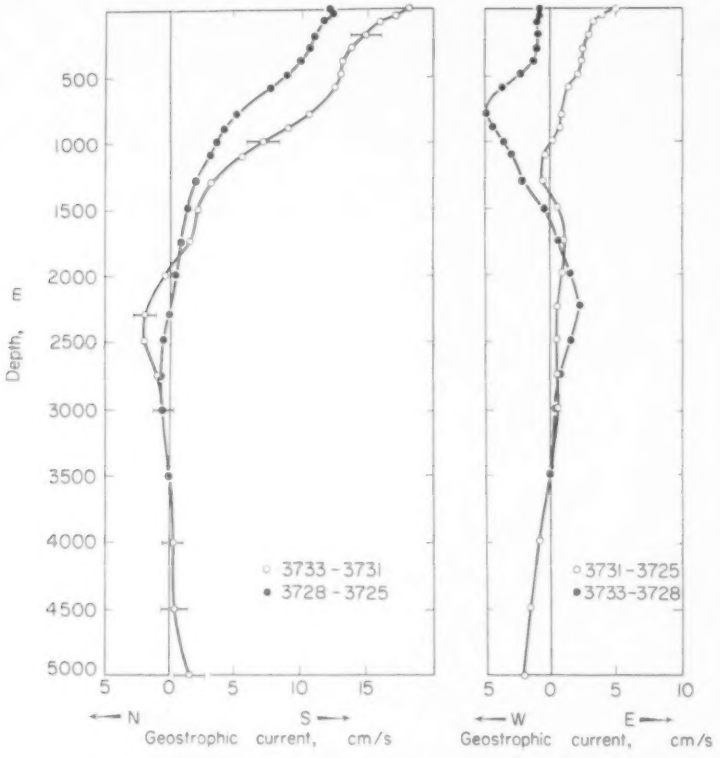

Fig. 6. Geostrophic currents, relative to 3500 m depth, from *Discovery II* Stations 3725, 3728, 3731 and 3733.

Fig. 6 shows the north-south and east-west geostrophic currents relative to 3500 m, calculated from each pair of stations. Estimates of the standard deviations of the computed relative currents at a number of depths have been added to one of the curves of Fig. 6. These estimates are based on the assumptions of independent standard deviations of $\pm 0{\cdot}015$°C in temperature and $\pm 0{\cdot}004$‰ in salinity, and negligible errors in depth and distance between stations. Similar limits apply to all four curves. The differences between the two members of each pair of curves are of the order to be expected, from the standard deviations, and the relative currents below the reference level of 3500 m are almost certainly not significant.

For comparison with the float results, the mean north-south and east-west geostrophic currents were computed, and their components in the direction in which the floats moved in the first period (taken as 143°T) were added. The zero of the curve of current against depth found in this was displaced so as to give the best fit with the observed movements of floats 7, 2, D, B and 11, and the result is shown in Fig. 7. The currents indicated by floats 2 and D do not fit the curve very well. It is difficult to say to what extent these discrepancies are due to errors in the shape of

the geostrophic current curve. Float D slowed down soon after the velocity plotted in Fig. 7 was measured, and it is possible that some slowing-down had already occurred in the eleven days between the occupation of the hydrographic stations and the measurement of velocity.

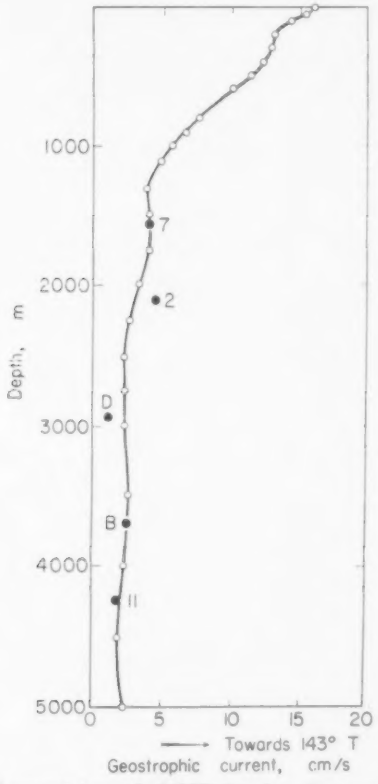

Fig. 7. Mean component of current towards 143°T, fitted to the currents observed with floats 7, 2, D, B and 11.

Helland-Hansen made a hydrographic section extending from the continental slope to 20°W during 18th-24th June (Fig. 1). The geostrophic currents across the deepest part of the section, calculated relative to the 5000 m depth between alternate pairs of stations, are presented in Fig. 8. They are more than ten times as strong as the mean current across the whole section, plotted in the same diagram on ten times the scale. The observed variations along the section in the deep water appear to be significant, and are comparable in magnitude with the variations observed directly on a small scale. Fig. 9 shows the calculated currents, relative to the 3000 m level, between the pairs of stations in the section nearest to the area of the direct measurements. These stations were occupied on 21st June, long after the south-eastward movement in the deep water had stopped. Apparently, though, the southward flow in the upper 1000 m was still nearly as strong as that computed earlier (Fig. 6).

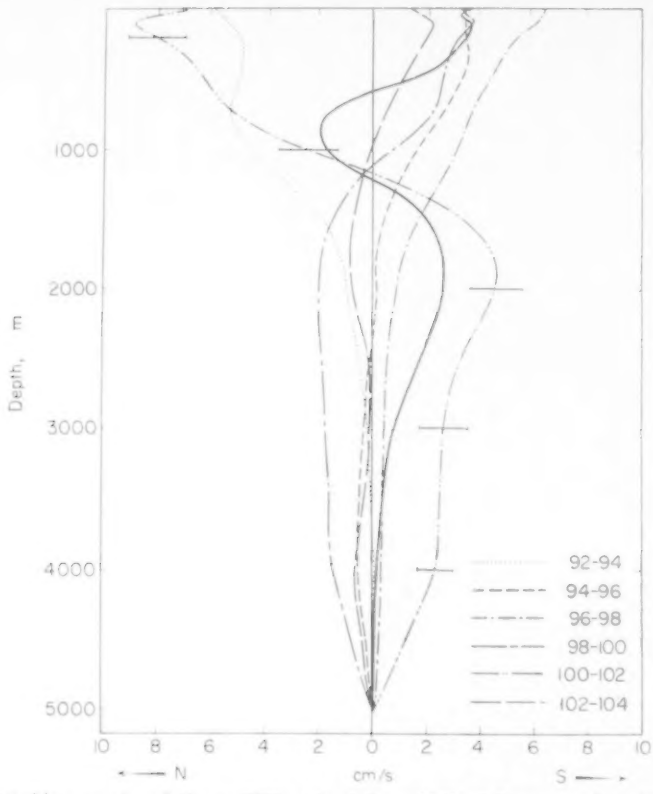

Fig. 8. Geostrophic currents, relative to 5000 m depth from *Helland-Hansen* stations 92-104. For the dashed curves, the current scale is in cm/sec. The solid curve represents the mean current profile across the whole section, for which the scale should be read in mm/sec.

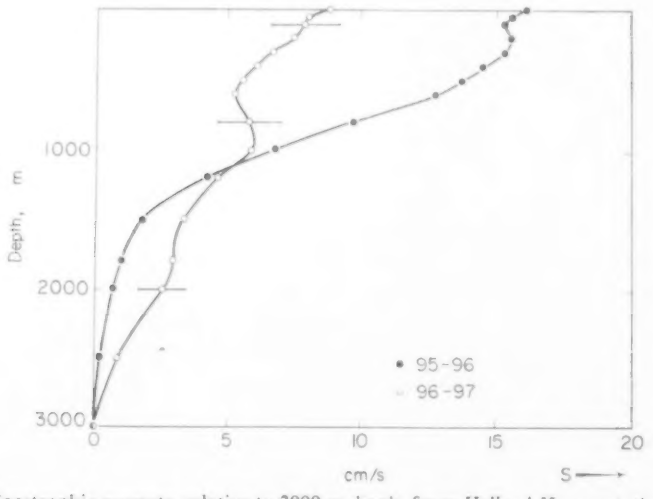

Fig. 9. Geostrophic currents, relative to 3000 m depth from *Helland-Hansen* stations 95-97.

A 'square' of hydrographic stations occupied by *Helland-Hansen* during 10th-12th July showed a changed situation in the upper layers. Currents relative to 5000 m depth were computed using stations 145,147, 151 and 153 (see Fig. 2) and the N-S and E-W components are shown in Fig. 10. At that time, float D, at 2940 m depth, was moving south-west at 1 cm/sec. The profiles indicate a predominantly westward movement in the upper 1000 m, and suggest that a weak westward flow may extend to the bottom, but it is not very significant.

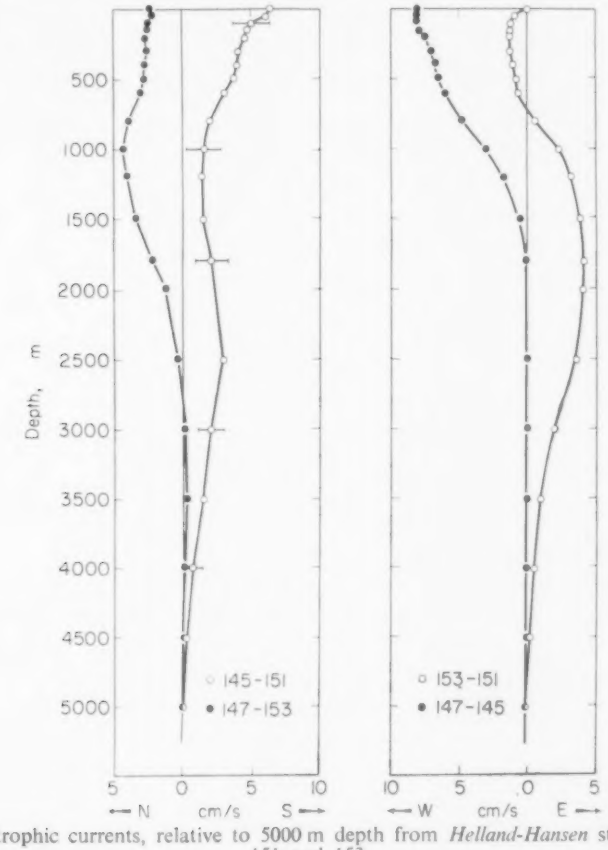

Fig. 10. Geostrophic currents, relative to 5000 m depth from *Helland-Hansen* stations 145, 147, 151 and 153.

Two lines of closely-spaced stations were worked by *Helland-Hansen* during 26th-29th June, approximately at right angles to the tracks of floats 3, 5 ,6 and D. Samples were taken at 500 m depth intervals from 200 m to 5000 m, in an attempt to study further the remarkable differences of velocity observed. The standard deviations of the computed currents were larger than in previous cases, due to the smaller horizontal spacing of the stations, and the calculated differences of current are not significant. When the geostrophic currents across the two sections are averaged, there is still no indication of agreement between the observed currents and the geostrophic currents relative to 4500 m. It appears likely that there was an appreciable westward velocity below floats 5 and 6, at depths as great as 5000 m.

ESTIMATES OF ACCELERATION AND FRICTION

The equations of horizontal motion in Cartesian co-ordinates can be written

$$\frac{Du}{Dt} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + F_x \tag{1}$$

$$\frac{Dv}{Dt} + fu = -\frac{1}{\rho}\frac{\partial p}{\partial y} + F_y \tag{2}$$

where u, v are velocities in the x, y directions respectively, p is the pressure, ρ the density, $f \equiv 2\omega \sin\phi$ is the Coriolis parameter, $\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y} + w\frac{\partial}{dz}$ is the total differential coefficient following a particle of fluid, and F_x and F_y are friction terms. The equations from which the geostrophic currents are computed are derived from (1) and (2) by assuming that the first and last terms are negligibly small.

Since the floats are presumed to follow the motion of the water in a horizontal plane, the D/Dt terms can be estimated from the observed tracks. Uncertainties of position-fixing, and long intervals between fixes, make it difficult to evaluate these accelerations in most cases. However, it can be seen from Fig. 4 that between 10th and 25th June, floats 3 and D accelerated significantly to the westward. The magnitude of this acceleration is approximately 7×10^{-7} cm sec^{-2}. This is small compared with fu and fv, which are of the order of 10^{-4} cm sec^{-2} in mid-latitudes.

The friction terms F_x and F_y cannot be determined so easily. Friction due to vertical shear, in the form $\frac{A_z}{\rho}\frac{\partial^2 u}{\partial z^2}$, was estimated from the five floats used in the first part of the cruise. A_z is a coefficient of eddy viscosity, for which values ranging between zero and 7500 have been determined by various methods (SVERDRUP *et al.*, 1942, p. 482). Taking $A_z = 3000$ gm cm^{-1} sec^{-1}, the maximum value obtained for the vertical friction term was 3×10^{-6} cm sec^{-2}, which again is small compared with fu and fv.

THE MERIDIONAL VOLUME TRANSPORT

The meridional volume transport, during the first part of the cruise, was calculated by integrating the meridional component of the velocity shown in Fig. 7 with respect to depth from surface to bottom. The current was assumed to decreased linearly to zero between 5000 m and the bottom. The value obtained for the volume transport was 180 m^3/sec to the south, per metre width of the current.

In steady conditions, and with a constant depth of water, the meridional volume transport should be proportional to the convergence of the surface wind-drift layer (STOMMEL, 1956). This is a consequence of the variation of the Coriolis parameter with latitude, the relationship being

$$F(0) = \frac{\rho}{f}\frac{\partial f}{\partial y} Tv \tag{3}$$

where $F(0)$ is the surface convergence, Tv is the volume transport, and y is distance in the meridional direction. Using the above value of Tv, $F(0)$ is found to be $3{\cdot}2 \times 10^{-3}$ gm cm^{-2} sec^{-1}. The climatological mean value of the convergence in this area, for July, is however no more than about 2×10^{-5} gm cm^{-2} sec^{-1} (MONTGOMERY, 1936). A convergence as large as that calculated from the volume

Vol 6 1959

transport would be produced by a 25-knot wind that decreased to zero in a distance of 100 km. Such a steep gradient seems unlikely, especially when averaged over a week or two. It is concluded that the observed meridional transport was probably not in balance with the surface convergence. A slope of the abyssal plain upwards towards the south could reduce the discrepancy, but the observed slope is in the opposite direction.

CONCLUSIONS

The two main results of the work are that currents in the deep water can vary, in the range 0-5 cm/sec, over distances of a few tens of miles and over periods of a few weeks, and that a large integrated volume transport in the meridional direction can apparently exist for at least a week. These results suggest that much more extended measurements will be needed for comparison with steady-state models.

VERONIS and STOMMEL (1956) have given a mathematical analysis of the different types of motion to be expected in a stratified ocean subjected to variable wind stresses. Of the six wave motions discussed, only two appear to be associated with currents of the order of magnitude observed by means of the floats. These are both geostrophically-balanced wave motions, in which the wave can be propagated only in a westerly direction. Whilst one of these motions can have a period short enough to be of the order required by the observed time variations, neither can account successfully for the space variations, since their wavelengths are too great. ROBINSON and STOMMEL (1959) have shown that irregularities in bottom topography might cause appreciable variations of current over short distances, but it is doubtful whether the ridges shown in Fig. 2 are high enough to be effective. The movements of floats D and 5 were not significantly affected by the ridges passed over, towards the ends of their tracks.

The observed changes in deep water movements with time appear to be unrelated to any large fluctuations in the upper layers, though changes of the same magnitude as the deep-water variations could have occurred and would not have been detected. The space variations suggest that the scatter often found in computed velocities across hydrographic sections may sometimes be real, and need not be entirely due to internal waves or to errors of measurement.

Acknowledgements—The measurements described above involved the efforts of many persons. The authors are particularly grateful for the co-operation of Professor H. MOSBY and his colleagues on board the M.S. *Helland-Hansen*, in making hydrographic stations in the most useful positions for interpreting the current measurements, and for permission to make use of the hydrographic data before publication. Of those on board the R.R.S. *Discovery II*, Captain J. D. GRAY and the deck officers were especially helpful in the unusual navigational work required by the current measurements. Thanks are due also to Dr. M. N. HILL and other members of the Geophysics Department of the University of Cambridge who gave assistance during the experiments. Among the authors' colleagues from the National Institute of Oceanography who took part in the cruise, particular mention must be made of Mr. H. CHARNOCK, Mr. J. CREASE and Dr. A. S. LAUGHTON, for their practical contributions to the current-measuring and water-sampling and for valuable discussions. The authors are indebted to the Chief Scientist and members of the staff of H.M. Underwater Detection Establishment, for assistance in obtaining and repairing transducers. One of us (B.V.H.) acknowledges assistance from UNESCO in the form of a Fellowship in Oceanography.

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Vol
6
1959

INSTRUMENTAL NOTES

Suggestion for eliminating pressure effects on protected reversing thermometers

(*Received* 22 *June* 1959)

It was pointed out recently (Folsom *et al.*, 1959) that present day reversing thermometers appear to be not entirely ' protected ' from the effects of external hydrostatic pressure, and that a correction for the error due to pressure should be considered whenever very precise measurements are to be made at great sea depths. Reasons were given for expecting an average modern protected thermometer to indicate temperatures too high by about 0·002 C for each 1000 m increase in depth ; and it was shown that certain instruments, those overfilled with mercury and those designed with too small an air volume below the cork, could deviate because of pressure by as much as 0·008°C per 1000 m depth change.

The rate of change of apparent temperature with depth varies with the depth ; it is greatest for large depths. Its numerical magnitude depends upon difficult-to-measure details of construction, and it would be difficult to compute for any completed thermometer. Rather elaborate equipment would be required to determine this correction in the laboratory ; however, it might be determined or compensated for at sea by making two successive casts with instruments transposed as to depth. The latter procedure would result also in a precise measurement of the temperature gradient ; it is perhaps only in gradient determinations in very deep water that individual reversing thermometers must be corrected for pressure effects.

It is apparent that some slight changes in the construction of the protected thermometers would diminish the pressure error by at least one order of magnitude, making it insignificant for all oceanographic purposes. In the present instrument, the sea pressure is able to increase the pressure in the small volume of air confined below the cork by an appreciable fraction, because the mercury displaced by the decrease in the whole lower portion of the external shell crowds into this small air space. Obviously, means should be sought for increasing the ratio of the volume confining air to the deformable volume. It is clear that a gas communication through the cork would provide a suitable means, because of the large space above the cork. A simple, open tube through the cork might be sufficient; and the pouring of mercury through this opening when the instrument was upset should not lead to serious new objections. Of course, the present brass fittings would have to be replaced by fittings resistant to mercury.

A much neater solution would be afforded should the maker pierce the present cork with a tube capped at the lower end, or filled with ordinary sintered glass of a pore-size allowing air to pass readily but permitting no mercury passage under the pressures involved here. It would appear that this would add only a few minutes more construction time and might even facilitate the overall manufacture and calibration since it would decrease the likelihood of a slow change in calibration due to diffusion of confined air through the cork.

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Acoustically monitored bottom coring

(*Received* 29 *June* 1959)

Abstract—An adaptation of the sound pinger designed by EDGERTON to position an underwater camera for bottom photography has been used to monitor bottom coring operations with a Stetson corer. It is found that the pinger portrays the whole procedure in considerable detail, including the triggering of the corer for free-fall at the bottom. Cores have been taken successfully in water depths of 1912 and 2635 fathoms in the Ionian Sea.

INTRODUCTION

IN OPERATING instruments near the ocean floor in deep water one of the most demanding difficulties has been positioning the gear properly with respect to the bottom. Acoustical monitoring devices have been suggested several times in the past, and a few such devices have come into use. An example is the Scripps Institution's ball breaker (ISAACS and MAXWELL, 1952). Another more recent example is the EDGERTON pinger for monitoring bottom photography (EDGERTON, in preparation). In this note tests are described of an adaptation of the EDGERTON pinger for monitoring deep coring operations.

THE EDGERTON PINGER

The Edgerton Pinger is an ADP crystal array similar to the U.S. Navy's UQN-1b echo-sounding transducer manufactured by the Edo Corporation. The principal difference between the two is that the pinger is mounted in a much lighter aluminium case without the air cavity which houses the inductor of the UQN-1b. The transducer is shock-excited through a carefully tuned auto-transformer from battery power supply, so that no cabling to the surface is needed. The pinger radiates very short pulses of sound (pings) at a precisely controlled rate which is approximately one per second. It has been used by Edgerton and others to position underwater cameras above the bottom. His technique is to attach the transducer to the camera so that it points downward; then the ping and its bottom echo are recorded on an echo sounder recorder as the camera is lowered toward the bottom. The difference in travel time between the directly transmitted ping and the bottom echo is a measure of the height of the camera above bottom.

This instrument was tested by EDGERTON, R. H. BACKUS and the writer on USCGC *Yamacraw* in August 1957 using the Precision Graphic Recorder (PGR) (KNOTT and HERSEY, 1956). These tests were successful in permitting close control of an underwater camera within a few feet of the bottom in water depths up to 1500 fm. It occurred to the writer at the time that here at last was a sound source of adequate power and resolution to perform a similar service for bottom coring.

THE BOTTOM CORING TESTS

During the current cruise of USNS *Chain* to the Mediterranean the first opportunity presented itself for testing the pinger as monitor of a coring operation. At lat. 36° 55′N and long. 19° 50·2′E in the Ionian Sea we found a flat bottom at 1912 fm (assuming a sound velocity of 800 fm/sec). The bottom echo was followed by several later echo sequences suggesting the presence of several reflecting layers within 50 ft below the bottom. Some of these were well within reach of the 13-ft Stetson corer we had available at the time. A sample recording of the echo sequence over this flat area is shown in Fig. 1.

PROCEDURE AND RESULTS

The Stetson corer was rigged in the usual manner with 30 ft of suspension cable coiled at the trigger mechanism (HVORSLEV and STETSON, 1946) but without its lead weights. Thirty feet above the releasing mechanism we attached the EDGERTON pinger by means of a short wire rope pendant clamped to the suspension cable. Upon lowering this assembly we found immediately that the directly transmitted ping, and its echoes from the top of the corer, the trigger weight, and the sea bottom were all clearly recorded (Fig. 2). In addition there was a more faintly recorded echo which had been reflected first from the surface and then from the bottom before reaching the echo sounder.

Vol 6 1959

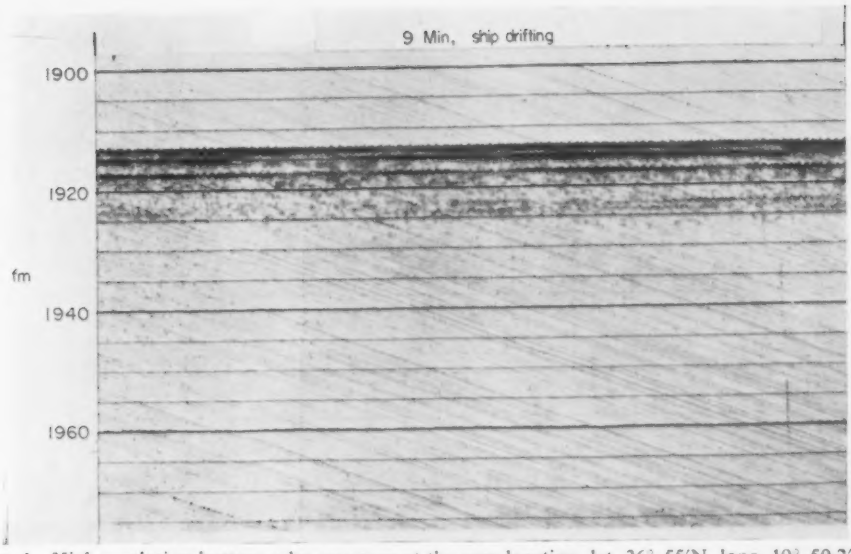

FIG. 1. High resolution bottom echo sequence at the core location, lat. 36°–55′N, long. 19°–50·2′E, from a drifting ship.

Vol.
6
959-60

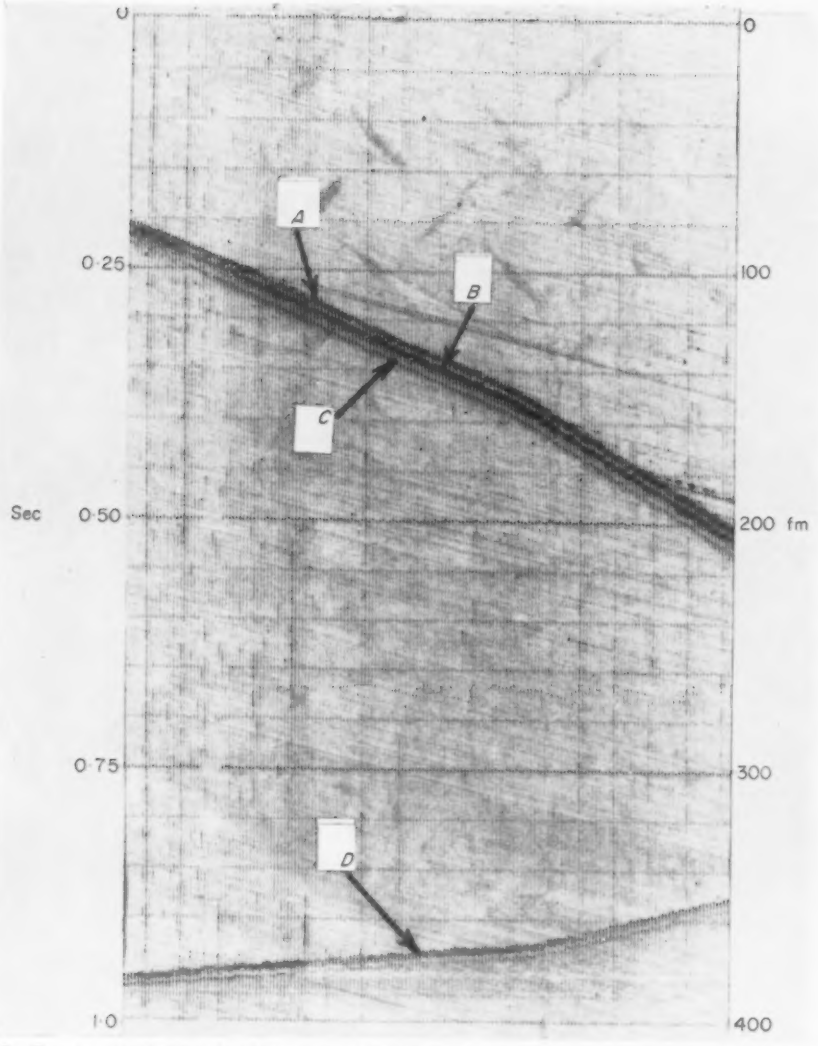

FIG. 2. Events recorded by the PGR when the corer and pinger are near the surface. A—the ping; *B*—the echo from the top of the corer; *C*—the echo from the trigger weight; and *D*—the bottom echo.

Vol
6
1959

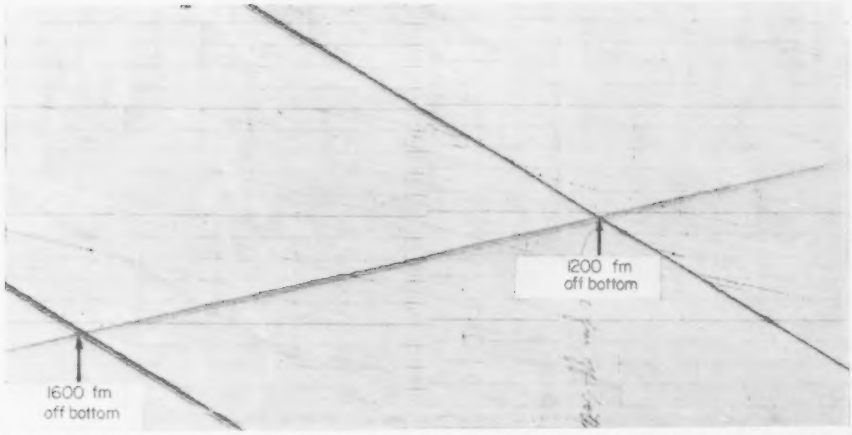

FIG. 3. The PGR record of part of the descent. Ping and bottom echo sequences crossed when the difference between their travel times was an integral number of seconds

Vol.
6
959-60

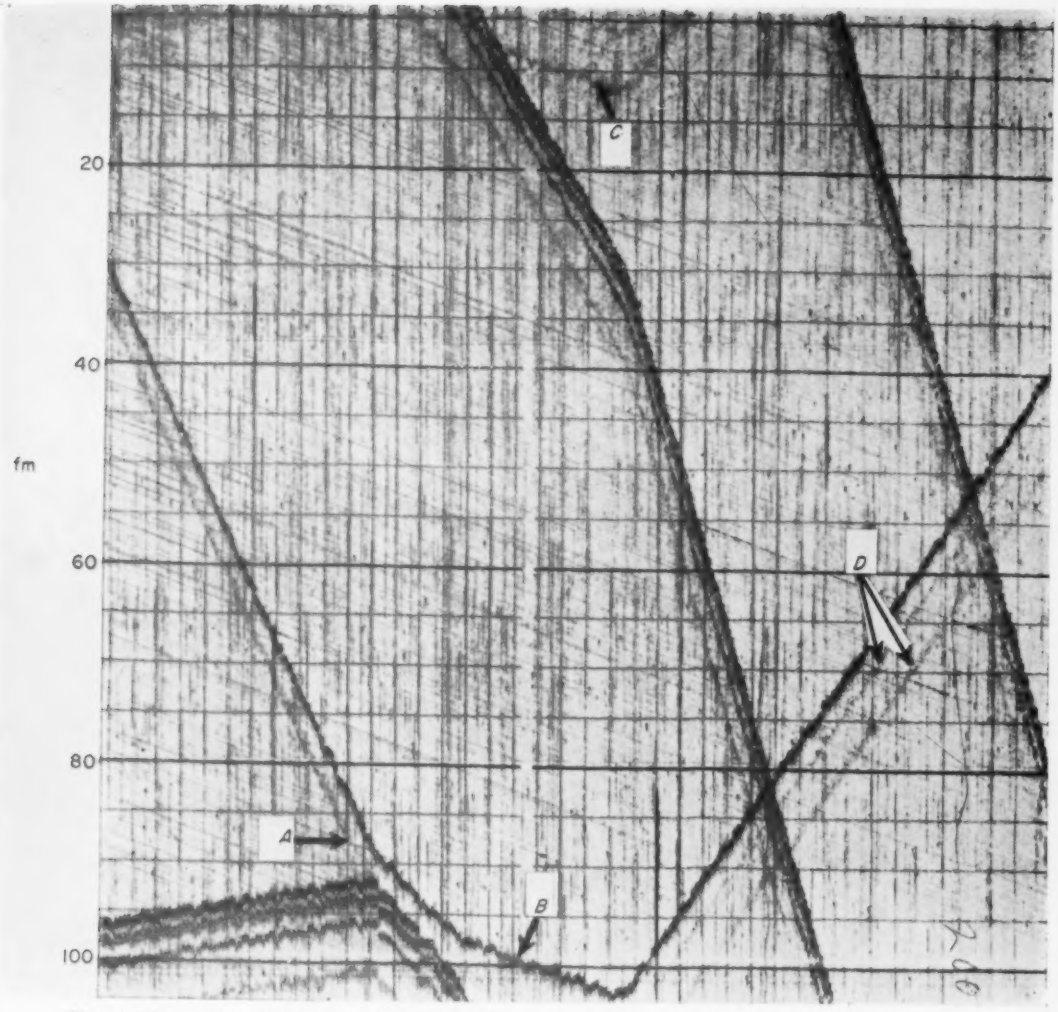

FIG. 4. Events recorded by the PGR when the corer is near the bottom. *A*—the echo of the top of the corer disappeared as the trigger released the corer for free-fall; *B*—the ascent commences; *C*—the echo of the corer appears as the corer is raised off bottom; *D*—during ascent echoes from both trigger mechanism and corer are recorded.

Vol
6
1959

Since the EDGERTON pinger pings once a second the PGR was operated with a one second sweep (400 fm scale) so as to record one ping and one bottom reflexion per sweep. Knowing the approximate depth of the pinger and the depth of water a precise measure of the difference in acoustic travel time between the ping and its own echo may easily be read. If the suspension cable had been vertical the exact height of the pinger above bottom could also be measured. In this lowering, wire angles were small (about 30° at most) and in any event the exact height is of little consequence; the principal objective in coring is to lower the corer deep enough so that the trigger weight operates satisfactorily without at the same time paying out extra wire to tangle on the bottom.

Thus knowing the water depth to be approximately 1900 fm the 'height' of the pinger and corer above bottom is easily deduced by recalling that the echo recorded on a given sweep must be that corresponding to the ping recorded about 4·75 sweeps previously. As the lowering proceeds the ping and echo sequences cross and re-cross at times when the travel times differ by an integral number of seconds, or multiples of 400 fm. Thus the progress of the lowering could be followed with assurance (Fig. 3).

When the height above bottom became less than 100 fm the PGR sweep was changed to 0·25 sec (100 fm scale). Then much more detail was apparent (Fig. 4). The bottom echo which had appeared as double on the slower sweep was now resolved into four clear echoes, and after increasing the gain of the recorder the echo from the top of the corer was to be seen. As the corer reached a height of 5 fm its echo disappeared because the tripping weight had just struck bottom releasing the corer (Fig. 4). The suspension cable was then raised and the echo of the corer was seen separating from the bottom echo. The corer echo shows that the corer was sixty feet from the pinger in its proper position after coring. A third echo, probably that of the trigger mechanism was discernible thirty feet from the pinger (Fig. 4).

DISCUSSION

Some speculation can be made about the maximum depths to which EDGERTON's present pinger can be used for coring. Since the first test described above a successful core has been taken in 2635 fm. In this water depth it was still possible to record the corer echo just before the tripping mechanism operated. Since it has proved to be the weakest sound of the sequence excepting only the echo from the lead tripping weight, one seems justified in supposing that the pinger is more than adequate for most ocean areas. Unfortunately it was not possible to obtain meaningful quantitative estimates of echo levels on this deeper lowering because of a bothersome electrical noise.

No preparations had been made for quantitative measurements during the first test, but judging from past experience with the PGR and the appearance of this recording the direct arrival and bottom echo were probably over 20 dB above the background noise throughout the test, while the other echoes were obviously fainter. With so favourable a signal-to-noise ratio in 1900 fm, useful operation should be possible in at least 3800 fm, since the additional transmission loss should be of the order of 14 dB (6 db spreading loss for doubling the distance and about 8 dB attenuation for the additional 3800 fm travel path). It appears to the writer that the signal-to-noise estimates above are conservative enough that the EDGERTON pinger may well be loud enough to be useful for coring in all ocean depths, given expectably quiet receiving conditions at the echo sounder.

ACKNOWLEDGMENTS

The author is principally indebted to Prof. HAROLD EDGERTON for his help and kindness in advising about his design and supplying auto transformers for the adaptation of his pinger to such purposes as reported here. Mr. ROBERT COTELL built the adaptations, and Mr. THOMAS STETSON and Dr. RICHARD PRATT rigged the corer. I am much indebted to Capt. OLIVEY and Boatswain LADZENSKI of the USNS *Chain* for valuable advice and assistance in the coring operations. This work was done under Contract Nonr-1367 with the Office of Naval Research, U.S. Navy Department.

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Contribution No. 1052 *of the Woods Hole Oceanographic Institution. This work was performed under Contract Nonr-*1367 *with the Office of Naval Research, U.S. Navy Department.*

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Vol
6
1959

Micro-organisms as indicators of hydrological phenomena in seas and oceans—II

Investigation of the Deep Circulation of the Indian Ocean using microbiological methods

A. E. KRISS, M. N. LEBEDEVA and I. N. MITZKEVICH

Abstract—(1) The microbiological investigations of the Indian Ocean water masses at 61 stations along the sections between Africa and the Antarctic, the Antarctic and Asia, showed that Antarctic waters are characterized by low concentrations of heterotrophs (which assimilate only easily accessible forms of organic substance), whereas the equatorial-tropical waters are characterized by high concentrations of these micro-organisms.

(2) The distribution of heterotrophs densities and individual bacterial species clearly indicate southerly currents carrying water from the tropical Indian Ocean south to the Antarctic coast and current flowing northward as far as the equator and the northern tropical zone.

(3) The microbiological data indicates a more complicated deep circulation in the Indian Ocean and in the adjacent parts of the Antarctic Ocean than that suggested by SVERDRUP, *et al.*, (1956). The water masses flowing from the tropical zone to the Antarctic coast were observed in the eastern half at depths of 25–75 m, 350–600 m, 1500–2500 m, and 2500–4000 m. In the tropical zone Antarctic water was found at 30–70 m, 100–300 m, 2500–3000 m, 3000–3500 m and in the equatorial zone at 1100–1500 m, 2000–2500 m, 2500–3000 m, 3000–3500 m.

As reported elsewhere (KRISS, 1959) microbiological methods make it possible to detect even slight differences in the stratification of the water masses. Additional data for the hydrological structure of the Indian Ocean were obtained as the result of extensive investigations on the distribution of heterotrophic micro-organisms during the oceanographical expedition on the *Ob* in 1956–57.

These investigations, unlike earlier ones (GAZERT, 1912, PIRIE, 1912), were made in the sections between Africa and the Antarctic, the Antarctic and Asia, as well as along the coast line of the Antarctic. The microbiological stations covered the area (Fig. 1) from the coasts of the Antarctic almost to the northern tropic in the Bay of Bengal.

At these stations 1117 samples were collected at standard hydrological depths from the surface to very deep layers. All samples were immediately assayed by culturing on membrane ultra filters (KRISS, *et al.*, 1952) in the specially equipped laboratory on the *Ob*. Forty millilitres of water were filtered. The microbial cells collected on the surface of the membrane filter were grown for 5–7 days at 18–22° in the Antarctic, and at 28–35° in tropical latitudes. The colonies were then counted, and the microbial forms were separated by their cultural characteristics and placed on agars, numbers of every type of colony being counted.

The numbers of heterotrophs at various depths of the Indian Ocean in different geographical zones are shown in Figs. 2–5. Zeros do not indicate complete sterility of the water but, as other investigations (KRISS, *et al.*, 1952) showed, only low concentrations of the heterotrophs at a given depth. Hence, they were not detected in the 40 ml water sample.

Near the Antarctic coasts (60–70 °S) and in the subantarctic zone heterotrophs in different layers (based on the number of colonies) were few in number, i.e., from the 40 ml water sample there were 0 to 10 colonies (Fig. 2). However, among the low counts of microbial cells belonging to strains which propagate on complex protein

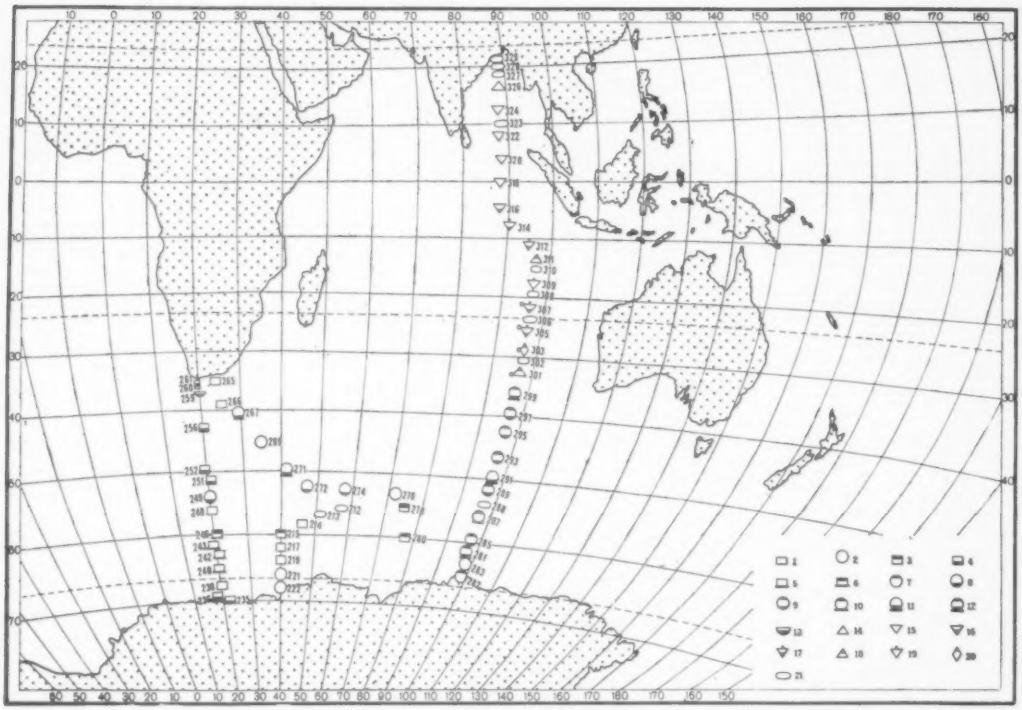

FIG. 1. Microbiological stations in the Indian Ocean. Stations where an increased number of micro-organisms in the water were discovered : 1 – at the depth 25 m; 2–25 and 75 m; 3–25 m and 450–550 m; 4–25 m and 1500–2500 m; 5–25 m and 3000–45,000 m; 6–25 m and 350–500 m and 300–4500 m; 7–25 m, 75 m and 400–550 m; 8–25 m, 75 m and 1500–2500 m; 9–25 m, 75 m, 400–550 m and 1500–2500 m; 10–25 m, 75 m, 400–550 m and 3000–4500 m; 11–25 m, 75 m, 1500–2500 m and 3000–4500 m; 12–25 m, 57 m, 400–550 m, 1500–2500 m, 3000–4500 m; 13–75 m and 1500–2500 m. Stations in the tropical zone where only a few micro-organisms were observed in the water layers: 14 – at the depth 1000–1500 m; 15 – 2000–3500 m; 16 – 200–3500 m and 4000–4500 m; 17 – 2000–3500 m, 4000–4500 m and 200–300 m; 18 – 2000–350 m and 1000–2000 m; 19 – 200–300 m and 2000–3500 m; 20 – 200–350 m, 1000–1500 m and 2000–3500 m; 21 – mud, coastal or suspended stations.

media there were cases when the concentration of heterotrophs was high. Hundreds and sometimes thousands of colonies developed on the filter surface from water samples collected at several stations from both the surface and deeper layers (Fig. 2).

Away from the antarctic and subantarctic regions near the tropical part of the Indian Ocean, the concentration of heterotrophs in the whole water mass increased considerably (Figs. 3, 4) as shown by the decrease in the number of 'sterile' layers and in the greater number of colonies grown on membrane ultra filters. A sharper increase in the number of heterotrophs in the oceanic depths was observed in the section along the 97th and 98th longitudes than along longitude 20.

Vol 6 1959

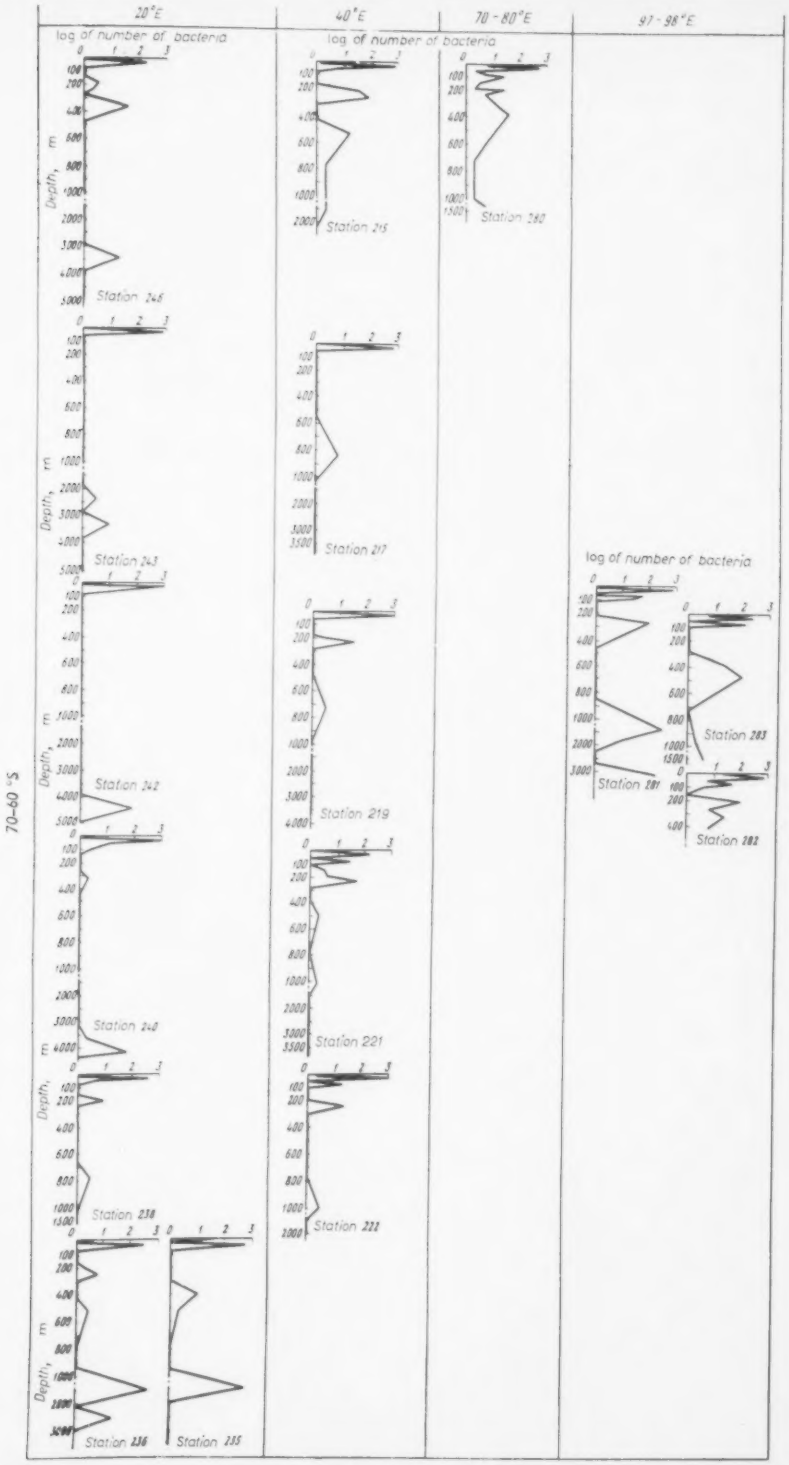

FIG. 2. Vertical distribution of heterotrophs in the Indian Ocean between 60 °S–70 °S. Number of bacteria per 40 ml volume of water.

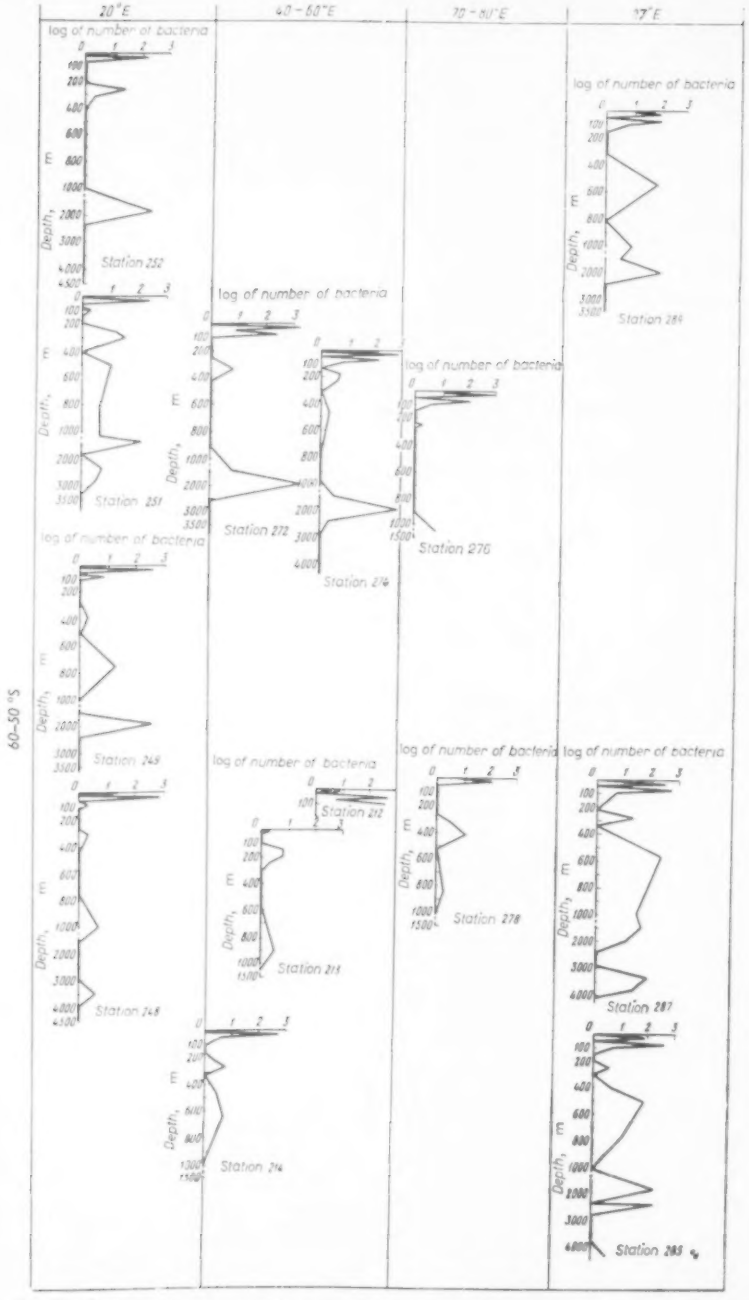

FIG. 3. Vertical distribution of heterotrophs in the Indian Ocean between 50 °S–60 °S. Number of bacteria per 40 ml volume of water.

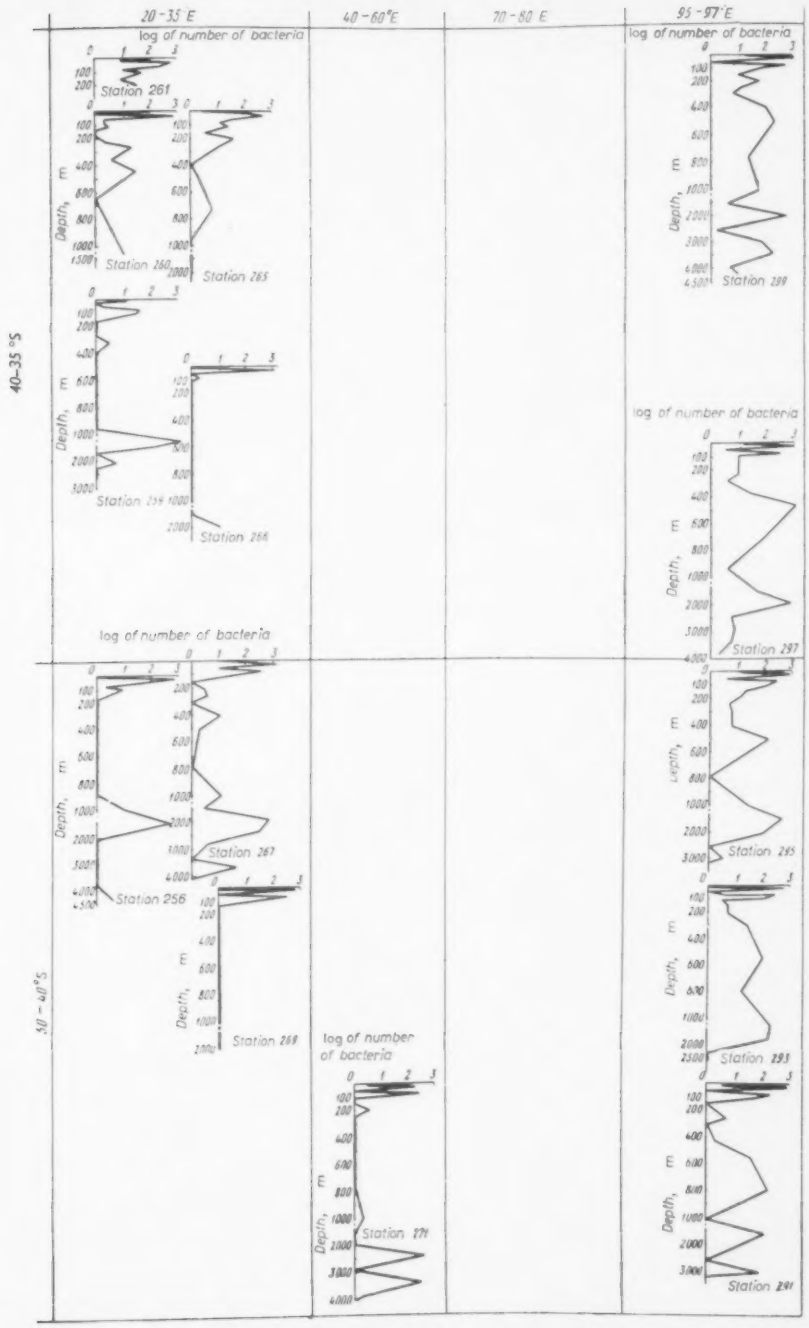

FIG. 4. Vertical distribution of heterotrophs in the Indian Ocean between 35 °S–50 °S. Number of bacteria per 40 ml volume of water.

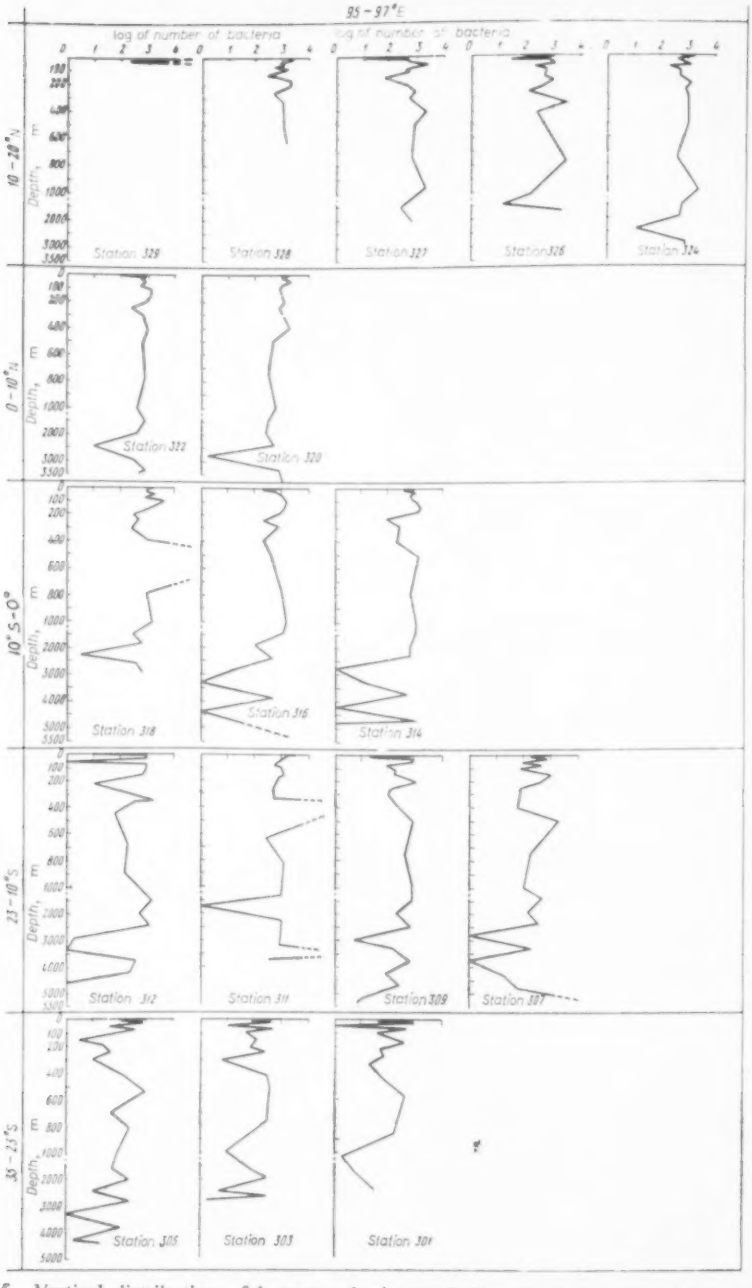

FIG. 5. Vertical distribution of heterotrophs in the Indian Ocean between 20 °N–35 °S. Number of bacteria per 40 ml volume of water

In contrast to the higher latitudes, where only single colonies or tens were observed, in the tropical and equatorial zones hundreds and thousands of colonies developed (Fig. 5) from 40 ml water samples from most layers from the surface to great depths (3000 m, 4000 m, 5000 m). The abundance of heterotrophs in tropical waters is, at first sight, of only quantitative nature. This is shown by the scarcity, as to the cultural characteristics of colonies, grown on filters. One is under the impression that waters rich in heterotrophs contain only a few species. However, it is not improbable that one or even a few quickly growing microbial forms cover the entire filter, suppressing forms with a lower growing capacity. In the subantarctic and antarctic regions, from water samples with few heterotrophs, a greater variety grew on the filters as a rule than in samples with a greater number.

From preliminary data, the diversity of species may be considerable in the microbial population of the tropical Indian Ocean and comparable to oceanic regions in higher latitudes. In 1058 cultures of heterotrophs from the Indian Ocean, the most numerous group – more than 700 cultures – were non-sporogenous mobile and immobile rods. The next largest group of about 200 strains, is represented by polymorphic cells of peculiar shape, probably belonging to mycobacteria. The other systematic groups of micro-organisms are distributed as follows : cocci – 100 cultures, sporogeneous rods – 40, yeasts – 11, Actinomyces – 5, Fungi – 3. Thus, the predominant forms utilizing easily assimilated organic matter, in the Indian Ocean as well as in other oceanic regions investigated, are non-sporogeneous rods.

It appears that in low latitudes of the Indian Ocean there is more organic matter easily assimilated by micro-organisms than in other latitudes. Micro-organisms, on account of their selectivity in assimilating organic substances, may serve as reliable indicators of qualitative differences in the organic composition of sea-water. Consequently, they are better indicators of slight differences in water masses than conventional chemical methods which indicate only the total organic matter in sea-water. Presumably, the enrichment of tropical waters with easily assimilated organic matter occurs not only at the expense of the plant and animal plankton. It is quite possible that numerous islands in the Coral Sea with very high concentration of life activity and extending over enormous areas of the tropical Pacific Ocean and areas bordering on the Indian Ocean, account to a considerable extent for the abundance of organic material still in early stages of decomposition and not transformed to humus in almost all layers of tropical waters. From these sources such organic matter, an ideal substrate for micro-organisms, is transported by the equatorial currents along the tropical zone, as well as by surface and deep currents which take them to higher latitudes.

The microbiological investigations clearly reveal the penetration of deep water from tropical regions into subtropical, subantarctic and even antarctic regions. The increase in the number of layers with ' zero ' content of heterotrophs and the decrease in the number of heterotrophs in the southern Indian Ocean clearly indicates the gradual dilution of the equatorial-tropical currents as they flow south.

There are, however, strong currents of equatorial-tropical water, which can be identified by the abundance of organic matter easily assimilated by micro-organisms, which penetrate deep into subantarctic and antarctic areas. A high concentration of heterotrophs near 25 m are seen at almost all stations from the subtropical zone to the Antarctic coast (Figs. 2, 3, 4, 5). Somewhat deeper, i.e. about 75 m, higher concentrations of heterotrophs occur along the entire 97th longitude from 22 °S to

64 °S and at some stations in these latitudes in the western Indian Ocean. Judging from the number of heterotrophs, the 350–600 m layer in the eastern Indian Ocean from 35 °S south is without doubt of tropical origin. This layer was also identified at some stations west of the section of the mouth of the Ganges to the Antarctic, but it is sometimes shallower at depths of 200 m, 250 m, 300 m as shown by the comparatively larger number of heterotrophs at these levels.

Of particular interest are the waters rich in heterotrophs at great depths in the Indian Ocean (1500–2500 m and 2500–4000 m). The numbers of heterotrophs differ little from those in the deep waters of the equatorial-tropical Indian Ocean and are doubtless of this origin. The microbiological indicators make it possible to follow the penetration of these waters as far as 60 °S, and in some cases still closer to the Antarctic.

It is easily seen that the deep-water currents along the 97th longitude penetrate farther into the subantarctic and antarctic than those in the western Indian Ocean. Along this longitude the increase of heterotrophs throughout the entire water column is more pronounced from the Antarctic towards the tropics. At the stations along the 97th longitude between 60° and 70 °S, there were fewer barren layers and the numbers of heterotrophs were somewhat higher compared with those along 40th and 20th longitudes (Table 1). Apparently, this may be attributed to the proximity

Table 1. *Distribution of heterotrophs at various latitudes in the eastern and western halves of the Indian Ocean.*

Latitudes	*Longitudes* 80°–90°–100°E					*Longitudes* 20°–40°–60°E				
	Number of samples investigated	*Number of colonies on filters*				*Number of samples investigated*	*Number of colonies on filters*			
		0	*below* 10	*below* 100	*over* 100		0	*below* 10	*below* 100	*over* 100
21 – 0 N	115	0·87	1·74	3·48	93·9					
0°–23°S	154	5·8	5·2	10·4	78·6					
23°–35°S	58		17·2	36·2	46·6					
35°–50°S	97	11·3	35·2	22·7	30·8	157	33·7	36·3	17·3	12·7
50°–60°S	106	33·2	29·2	20·8	16·1	157	46·5	38·2	4·46	10·8
60°–70°S	45	35·6	24·4	20	20	228	53·5	32·1	8·3	6·1

The figures express the relationship between the water samples and the corresponding number of colonies and the total number of the samples studied, (%).

of the eastern Indian Ocean to sources of organic matter easily assimilated by microorganisms, (i.e. to the numerous densely populated islands in the equatorial and tropical Pacific Ocean).

The microbiological data also indicate that in the deep Indian Ocean there are currents flowing from high latitudes towards lower ones and these are characterized by very low concentrations of heterotrophs. In some cases their concentration is so small that no colonies developed on a filter culture from a 40 ml sample. Thus, these waters do not differ markedly from main water mass in the antarctic and subantarctic regions of the Indian Ocean.

Waters with few heterotrophs were identified at depths of 1500–3500 m, 3500–4500 m at all microbiological stations in the equatorial-tropical Indian Ocean. This is evidence for the penetration of the antarctic waters to the equator and even to the north

Vol 6 1959

tropical Indian Ocean. The consistency of ' drops ' at great depths seen in the curves of vertical distribution of heterotrophs in the section along the 97th longitude clearly indicates the existence of currents flowing from the southern Indian Ocean to the tropics.

It should be noted that the circulation of water with low concentrations of heterotrophic bacteria to the north from the Antarctic and southern Indian Ocean also takes place in the layers between 30–70 m and 100–300 m. However, the flow above the zone of the subtropical convergence is considerably weaker, penetrating in narrow tongues only to the southern tropic and to 10 °S.

The movement of water masses in the Indian Ocean in opposite directions is also evident from the distribution of individual bacterial species. Though this question will be specially discussed later it is worth mentioning that microbial forms widely distributed in the equatorial-tropical zone were dominant in water layers spreading from the tropical zone to the Antarctic, where waters with high concentrations of heterotrophs are typical. On the other hand, in deep water of the equatorial-tropical area where there are few micro-organisms, these are forms usually found in the Antarctic Ocean. A schematic hydrological structure of the Indian Ocean (Fig. 6a, b)

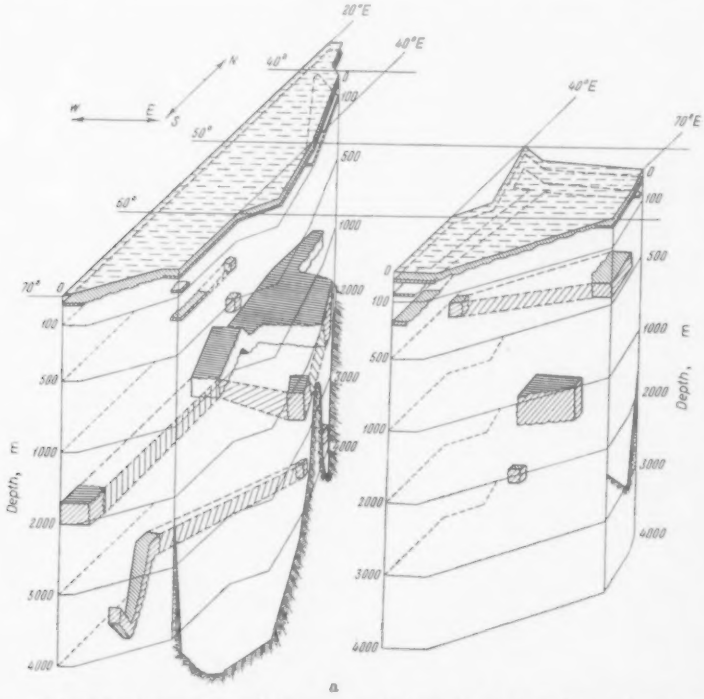

FIG. 6 (a). Hydrological structure of the western half of the Indian Ocean based on microbiological data. Shaded parts – layers of waters with a high concentration of bacteria.

based on the vertical distribution of heterotrophic micro-organisms reveals the existence of well defined deep-water currents flowing at various depths and in opposite directions. Some carry water from the tropics southwards to the Antarctic coast; others flow northwards to the equatorial and north-tropical zones.

From the vertical stratification of waters in the Indian Ocean suggested by SVERDRUP, *et al.*, (1946) its water mass is divided into the following layers : surface layer (0–100 m) ; Central or Equatorial Water (approximately 100–800 m), Intermediate Water (approximately down to 200 m); Deep and Bottom Waters (2000–4000 m).

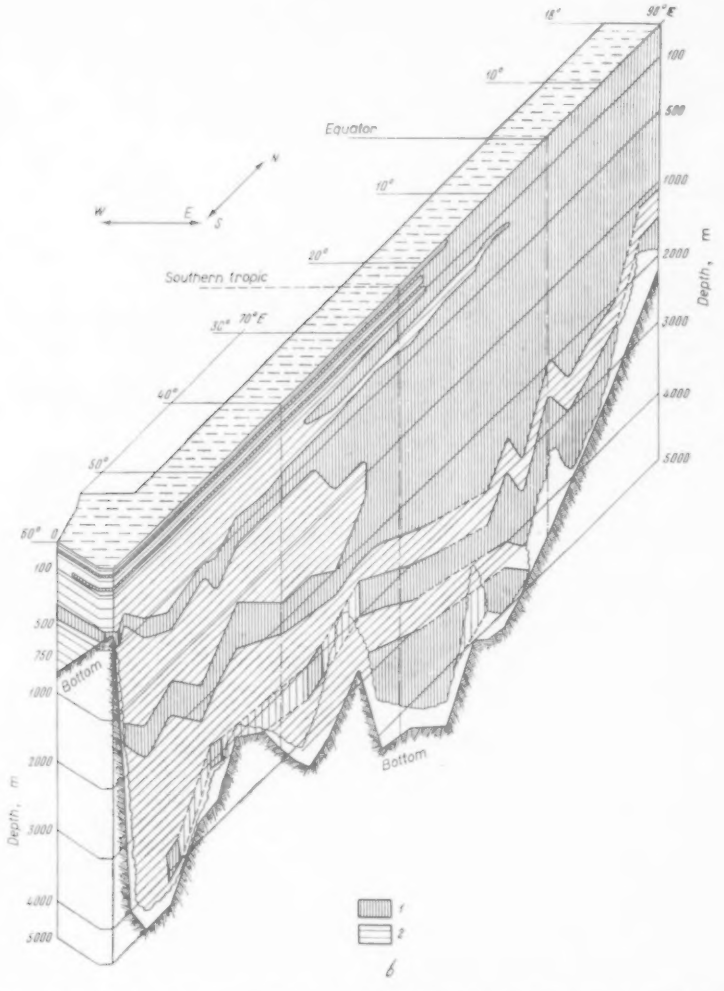

FIG. 6 (b). Hydrological structure of the eastern half of the Indian Ocean based on microbiological data.
1 – Water layers of equatorial-tropical origin with high concentrations of bacteria.
2 – Water layers of antarctic origin with low concentrations of bacteria.

The authors indicate that in the eastern Indian Ocean the intermediate layer is Antarctic water extending almost to 10 °S, the Bottom Water also flows northwards over the ocean floor in the equatorial area. The Deep Water flows in the opposite direction from north to south.

In the sector of the Antarctic Ocean where the conditional boundary for the Indian Ocean is the subtropical convergence, SVERDRUP, *et al.*, distinguished two areas : Subantarctic (between the subtropical and subantarctic convergence) and the more southern Antarctic area. In the Antarctic area below the surface layer (0–100 m) is a rather thin Intermediate layer. Beneath this down to depths of 1500 m is the Antarctic Circumpolar Water. The Bottom Water is still deeper.

The vertical stratification in the subantarctic area is somewhat more complicated. Below the surface layer (0–100 m) are thin layers of the subantarctic Upper Water below which are the Antarctic Intermediate Waters (approximately from 200–300 m to 1500 m). Below these flow Upper and Lower Deep Water layers and the Bottom Water. The Antarctic Intermediate Water and the Bottom Water flow northwards. Subantarctic Upper Water and Lower Deep Water flow southwards.

The new microbiological data showed the more complicated nature of the deep counter circulation in the water column of the Indian Ocean and in the adjacent Antarctic Ocean. The penetration of water masses from the tropical zone to the Antarctic coast occurs (in the eastern half) at depths of 25 m, 75 m, 350–600 m, 1500–2500 m and 2500–4000 m. A layer of water at 350–600 m lies below the Subantarctic Water, whereas the boundaries of the 2500–4000 m layer coincide approximately with those of Lower Deep Water (according to SVERDRUP's scheme, 1946). Not only do the Bottom and Antarctic Intermediate Waters flow northwards, according to SVERDRUP, *et al.*, but also water at 30–70 m and 100–300 m which occur in the Indian Ocean up to southern tropic and 10 °S. It should also be noted that the penetration of Antarctic waters to the equator and thence northwards takes place not only in the bottom layer. In the tropical zone, the Antarctic waters were clearly identified at 2500–3000 m and 3000–3500 m (Fig. 6b). In the equatorial zone these waters flow northwards at 2500–3000 m, 3000–3500 m and 2000–2500 m. In the Bay of Bengal, waters of the Antarctic origin were identified at the 1100–1500 m.

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The Ganges and Indus submarine canyons

P. J. D. HAYTER

(*Received* 15 *January* 1959)

Abstract—Details are shown of the topography of the sea floor near the heads of the Ganges and Indus Submarine Canyons.

INTRODUCTION

THE two canyons near the mouths of the Ganges and Indus Rivers (Fig. 1), known as Swatch of No Ground and The Swatch respectively, were included in hydrographic surveys carried out by the Pakistan Naval Surveying Ship P.N.S. *Zulfiquar* under the direction of Commander R. B. A. Hunt, O.B.E., R.A.N. Both are valuable submarine landmarks and were surveyed primarily for the benefit of shipping. However their oceanographical interest was appreciated and for this reason as much information as possible was gathered.

The surveys were carried out by conventional hydrographic methods and accurate fixing by horizontal sextant angles was obtained from a system of floating beacons whose positions were based on land triangulation. Lines of soundings, using echo sounding equipment, were run at approximately half-mile intervals in an East–West direction.

GANGES SUBMARINE CANYON (SWATCH OF NO GROUND)

The Ganges Submarine Canyon was surveyed in 1951 during the course of a survey of the approaches to the Pussur River of which the position is shown in Fig. 1. The details of the survey are reproduced in Fig. 2. The existence of the canyon has long been known and the Pussur River was indeed selected as a possible secondary port for East Pakistan because the Swatch of No Ground appeared to lead into it.

Unfortunately limitations of time and equipment prevented this survey from being completed and only about 25 miles of the head of the canyon is shown on the chartlet. From the few soundings that have been obtained on previous occasions it appears to extend a further 45 miles or so to the southward before debouching on to the floor of the Bay of Bengal. The portion which has been surveyed does however show the general formation of the canyon with its steep sides and comparatively level bottom.

No samples of the bottom were obtained in this canyon.

It has previously been suggested by DIETZ (1953) that the turbidity currents from this canyon extend southwards through the gently sloping Bay of Bengal and leave a channel with leveed sides in the otherwise flat bottom. A channel of this type has been identified at a position in the centre of the Bay of Bengal 900 miles south of the mouths of the Ganges. Similar channels have been observed in other parts of the world.

Vol
6
1959

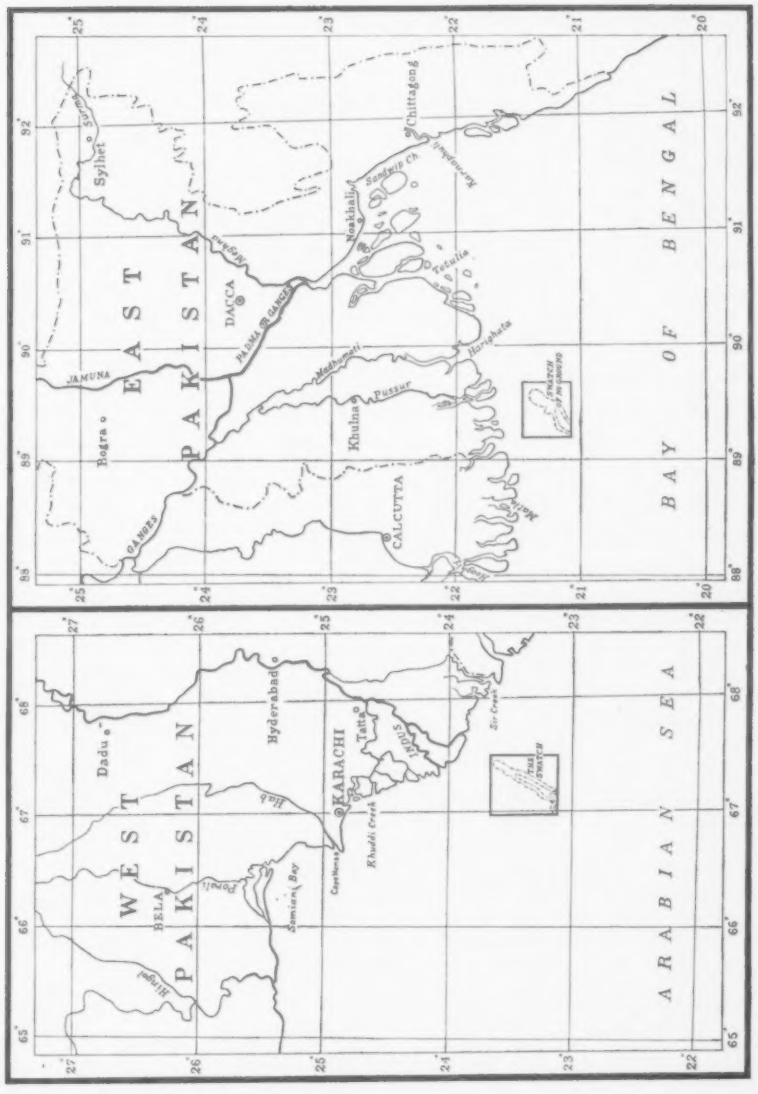

PAKISTAN
INDEX CHARTS SHOWING POSITIONS OF
THE SWATCH & SWATCH OF NO GROUND
EAST PAKISTAN
WEST PAKISTAN
BAY OF BENGAL
ARABIAN SEA
Sylhet
DACCA
JAMUNA
Bogra
Chittagong
Sandwip Ch.
Noakhali
Khulna
Rupsur
CALCUTTA
Hyderabad
Dadu
KARACHI
Tatta
Indus
Khadi Creek
Sonmiani Bay
Pasni
Hingol
Kolab
BELA
Swatch of No Ground
Trial Swatch

FIG. 1.

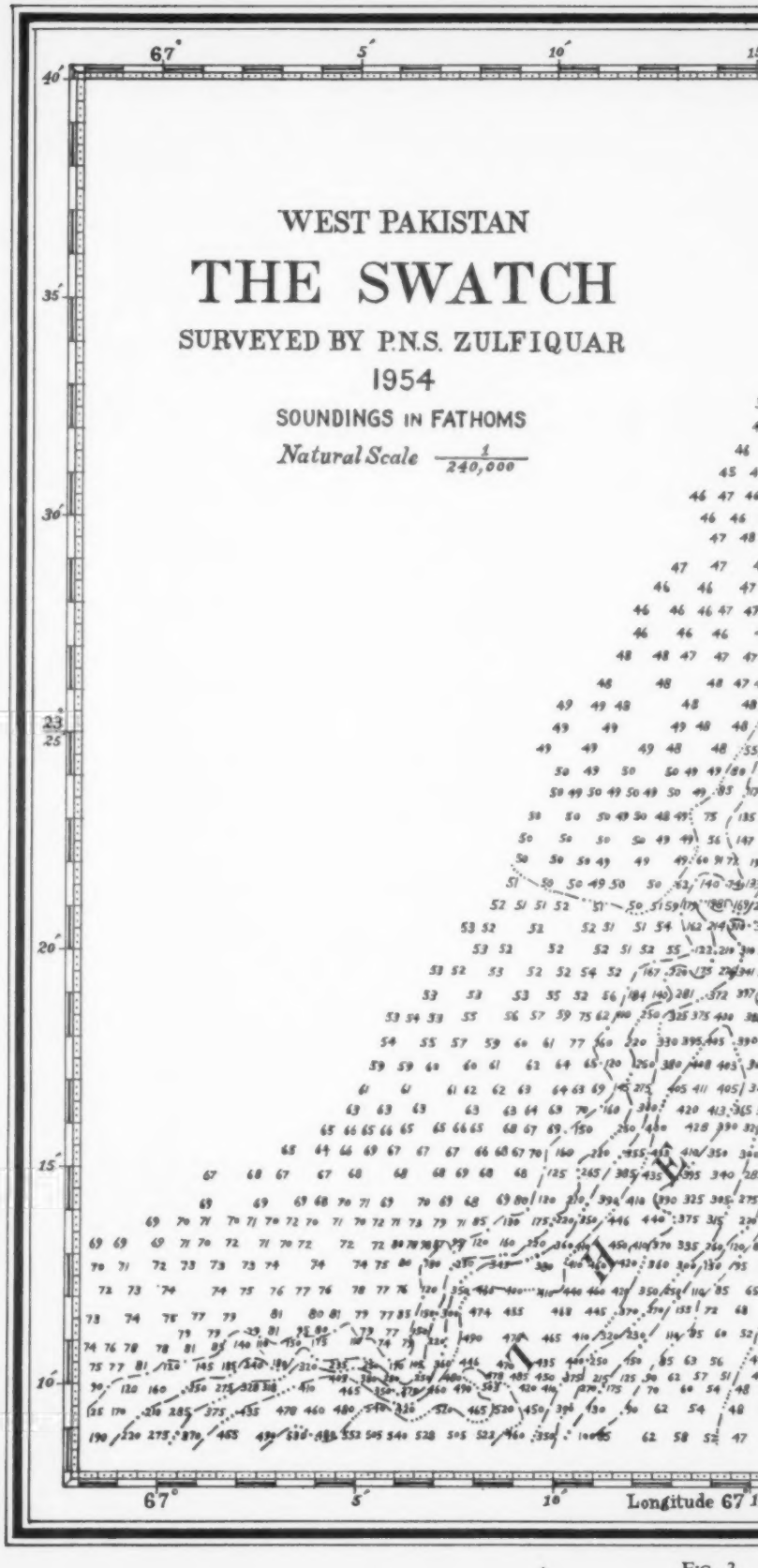
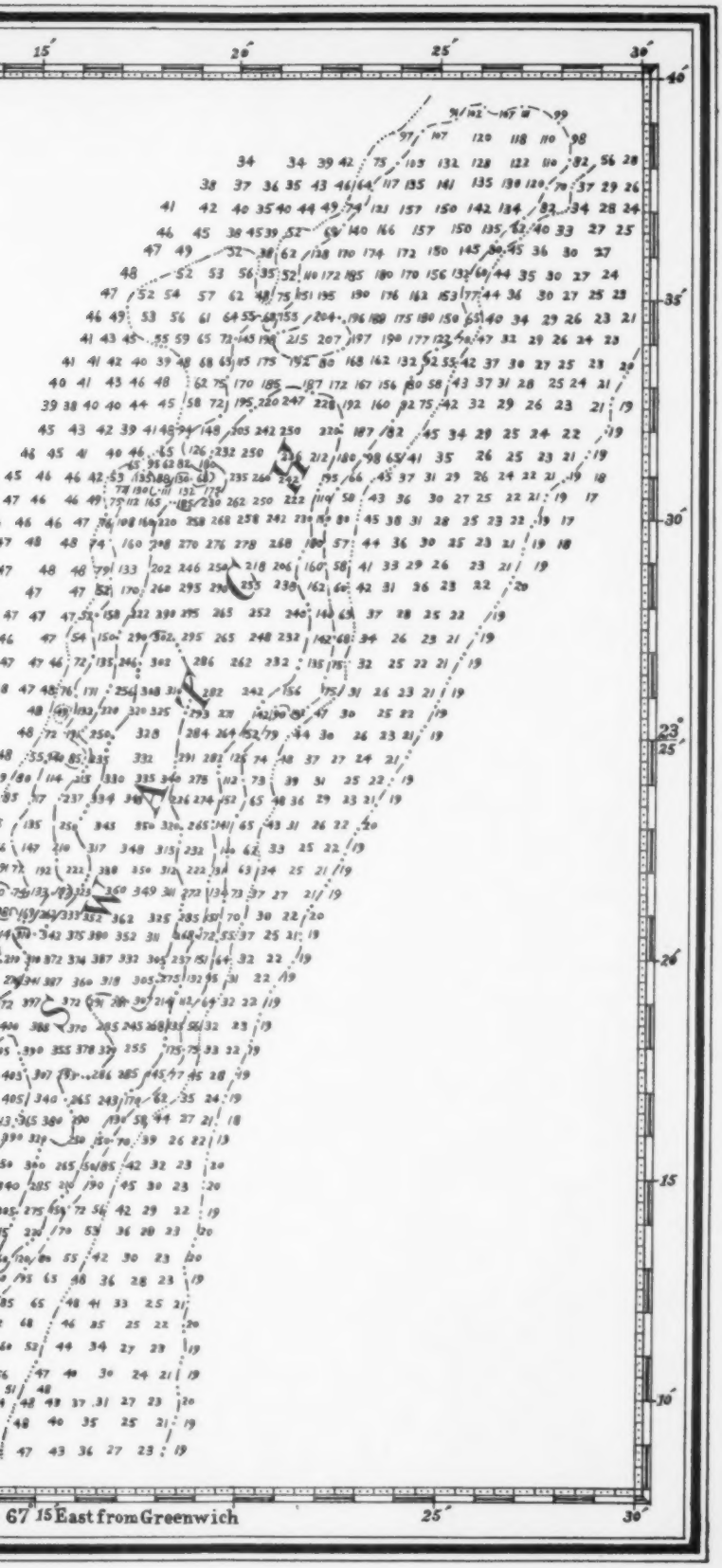

WEST PAKISTAN
THE SWATCH
SURVEYED BY P.N.S. ZULFIQUAR
1954
SOUNDINGS IN FATHOMS
Natural Scale 1/240,000
67°
5′
10′
40′
35′
30′
25′
20′
15′
10′
Longitude 67

FIG. 3.



. 3.

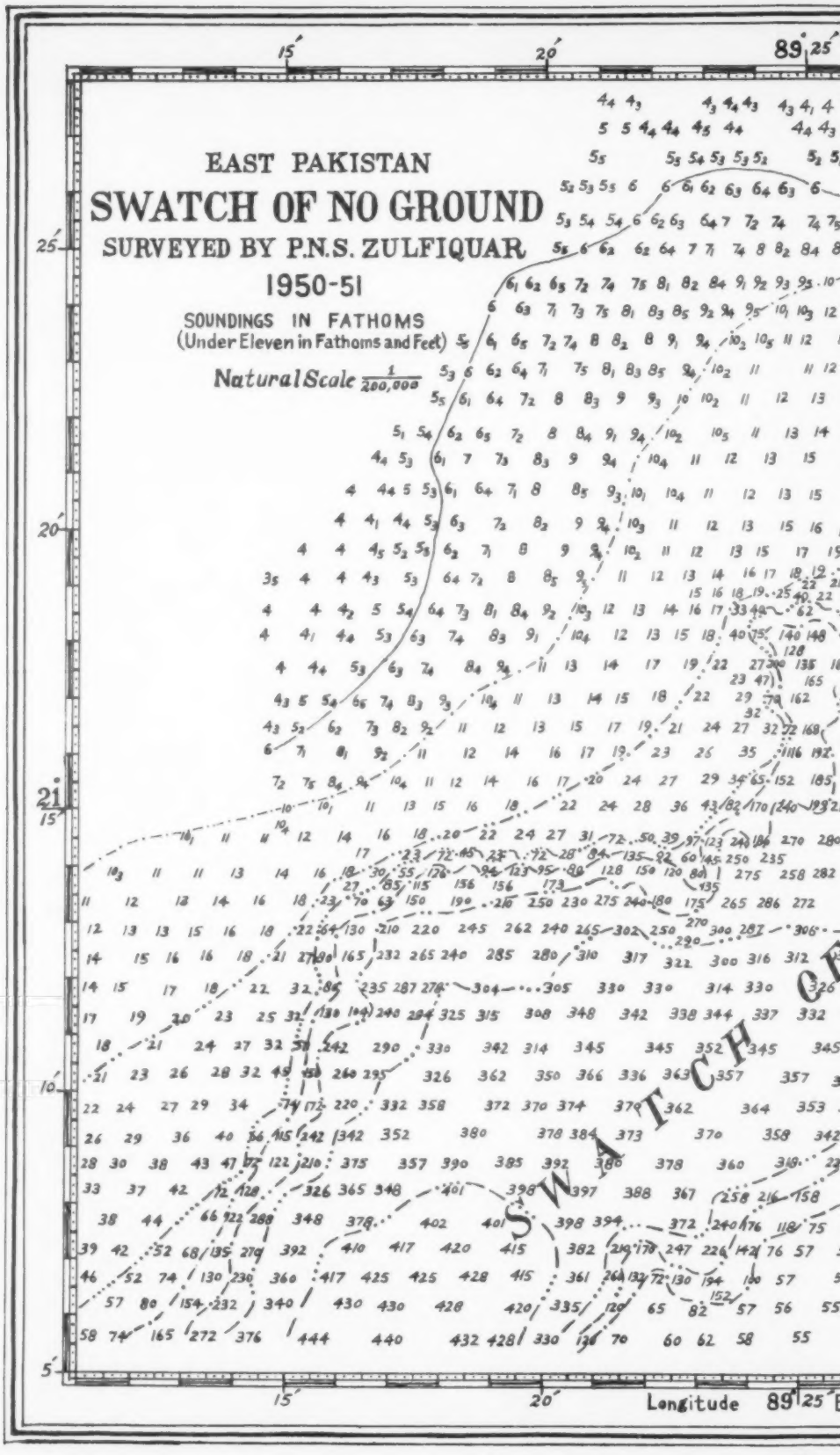

EAST PAKISTAN
SWATCH OF NO GROUND
SURVEYED BY P.N.S. ZULFIQUAR
1950-51
SOUNDINGS IN FATHOMS
(Under Eleven in Fathoms and Feet)
Natural Scale 1/200,000
SWATCH OF NO GROUND
15'
20'
89° 25'
25'
20'
21
15'
10'
5'
15'
20'
Longitude 89° 25'

FIG. 2

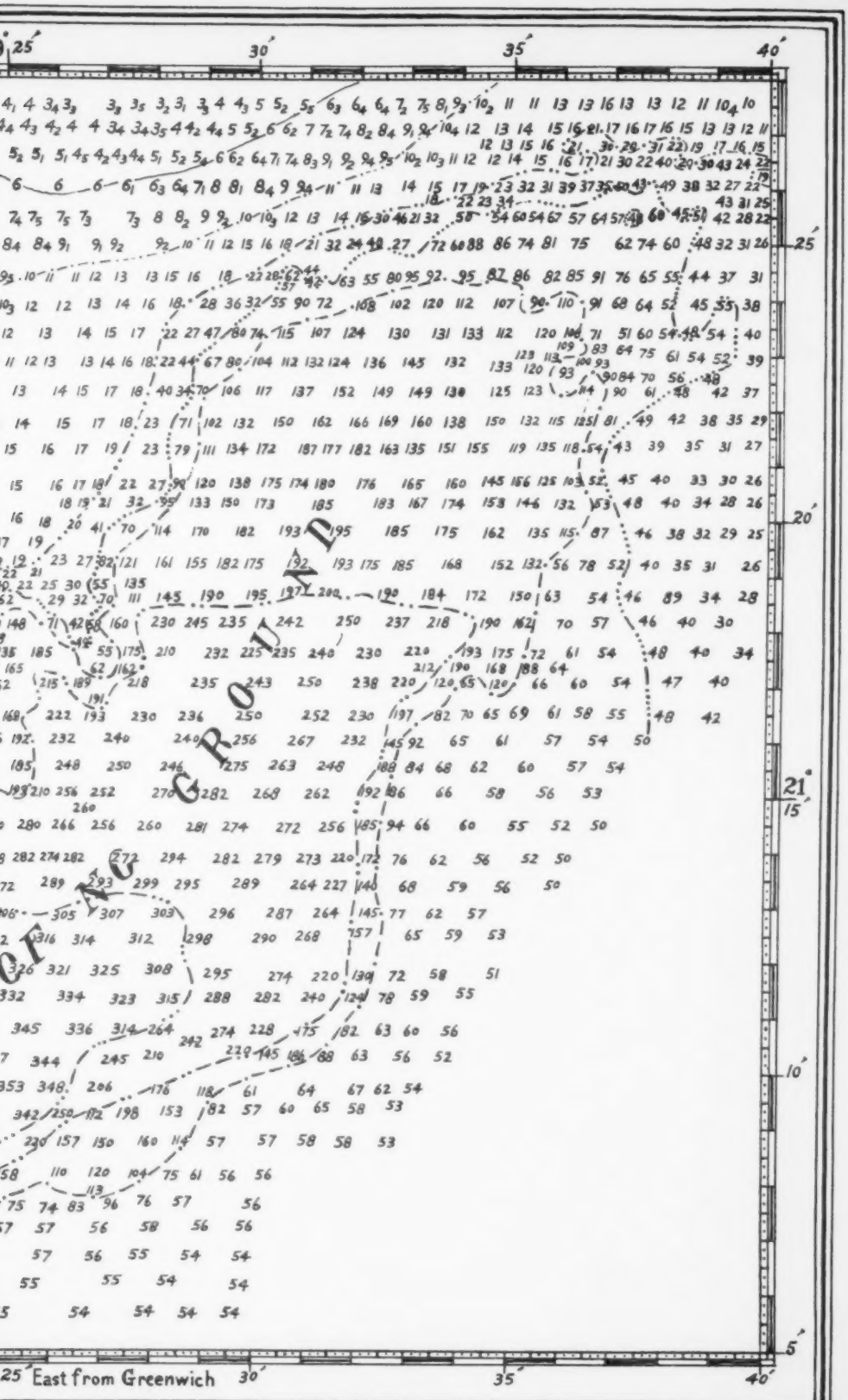


FIG. 2

4

INDUS SUBMARINE DELTA (THE SWATCH)

This was surveyed in 1954 during the survey of the approaches to the Indus delta on a scale of 1/75,000. A portion has been reproduced in Fig. 3.

The survey of this canyon is more complete although, because of limitations of equipment which did not allow soundings greater than about 600 fathoms to be obtained, it barely reaches the mouth. As far as could be ascertained the mouth, where it debouches on to the floor of the Arabian Sea without any marked fall, is bell-shaped and some 10 miles in width.

Transverse sections show that the slopes of the sides of the canyon are uniform ; and that across the bottom the depth is nearly constant. At the head of the canyon the longitudinal slope of the bottom is more gradual than further seaward ; the 100 fm line is about 12 miles and the 20 fm line 4 miles from the shore, giving a slope of 1:100. From the 100 fm to the 500 fm countour the slope is approximately uniform and has a value of about 1:80. The portion of the coast (near the mouth of the Kahr Creek) has encroached about 5 miles to seaward in recent years and where the coast is now, a sounding of 7 fms was shown in the survey of 1896.

Bottom samples in the canyon up to a depth of about 400 fms were all fine, dark grey, sandy mud, becoming slightly less fine in texture further to seaward. At several places along the top of the north-western slope, bottom samples of fine sand and small gravel were obtained. The Kori Great Bank to the south-east is also composed of fine sand, while the flats to the north-west are of mud of the same texture as the canyon itself.

The surface tidal streams appeared to run along the axis of the canyon and were slightly stronger than across the surrounding flats. There was normally during spring tides a slight overfall indicating the edge of the canyon. This has also been noticed over the edge of the continental shelf.

Naval Headquarters
Karachi, Pakistan

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Vol
6
1959

Les diatomées des dépôts de fond de la partie nord-ouest de l'Océan Pacifique

A. P. JOUSÉ

(*Received* 9 *April* 1959)

Résumé—Dans les dépôts du nord-ouest de l'Océan Pacifique les diatomées se trouvent en abondance et en bonne conservation. Dans la coupe verticale des carottes l'on observe des modifications concernant la composition des diatomées et aussi de leurs quantités.

Les diatomées des époques glaciales se distinguent par une plus froide composition que la flore actuelle du nord du Pacifique. Pour la composition des diatomées de la dernière époque interglaciaire sont caracteristiques les espèces de la partie sud-boréale contrairement aux diatomées nord-boréales de la flore de plankton actuelle au nord du Pacifique.

Vol.
6
959-60

DANS le présent article sont exposés les résultats fondamentaux des analyses des diatomées de quatre carottes obtenues au cours de la vingtième traversée du bateau de recherches *Vitiaz* par une équipe de l'Institut d'Océanologie de l'Académie des Sciences de l'URSS.

Les numéros et les coordonnées des stations étudiées, ainsi que leurs profondeurs et la longueur des carottes, sont indiqués dans la table No. 1.

Table 1.

Nos. des stations	*Situation géographique*	*Profondeur des stations (m)*	*Longueur de la carotte (cm)*	*Quantité des échantillons étudiés*
3342	54° 42·0′ n. lat., 164° 30·9′ long. de l'Est	4588	245	16
3325	52° 03′ n. lat., 166° 44·8′ long. de l'Est	4016	188	11
3359	51° 21·4′ n. lat., 172° 14·1′ long. de l'Est	4880	276	17
3378	50° 53·2′ n. lat., 162° 24·7′ long. de l'Est	5461	340	15

L'analyse des diatomées a été faite dans le but de connaître la stratigraphie des dépôts de fond dans les conditions paléogéographiques de sédimentation. L'abondance et la bonne conservation des diatomées dans les dépôts du nord-ouest du Pacifique nous ont permis d'appliquer cette méthode pour la solution des problèmes qui nous intéressent. Les diatomées, en s'accumulant en même temps que les dépôts terrigènes de l'océan, ne peuvent pas ne pas caractériser les conditions générales de la sédimentation.

Les diatomées sont surtout sensibles aux variations du climat, qui, à leur tour, influencent la température des masses d'eau de surface. Dans les dépôts du nord-ouest

du Pacifique on trouve surtout les espèces pelagiques répandues dans la couche d'eau supérieure ne dépassant pas cent mètres d'épaisseur. Leur présence dépend de la température de la couche supérieure de l'eau — ce qui est mis en évidence par l'analyse des espèces les plus typiques coincidant avec les limites des masses d'eau d'origine différente. Il en résulte qu'il est possible de juger des conditions paléoclimatiques de la sédimentation d'après la composition des diatomées fossiles. Les formes benthiques, surtout les espèces abyssales, se prètent difficilement à ce but.

Dans la couche supérieure des dépôts, la répartition des diatomées reflète fidèlement les conditions propres à leur diffusion dans le plankton de l'Océan d'aujourd'hui.

Dans les dépôts, la coupe verticale montre des modifications dans la flore des diatomées par rapport à celle de nos jours. Ces modifications touchent non seulement la composition spécifique, mais aussi le nombre des espèces et des individus.

Les carottes nous donnent des témoignages irréfutables des modifications des limites écologiques qui marquaient auparavent la présence des diatomées. A en juger d'après les carottes où les couches des dépôts, riches en diatomées, alternent avec celles où il n'y en a que peu, le développement des diatomées dans l'Océan Pacifique a été exposé à des variations.

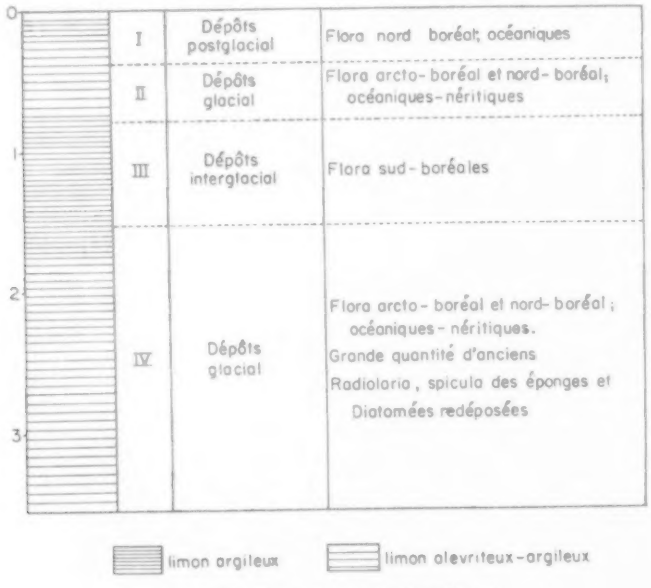

Fig. 1. Coupe schématique verticale des dépôts de la partie nord-ouest de l'Océan Pacifique.

L'analyse des carottes a démontré qu'en comparaison avec la couche supérieure des dépôts, les carottes ont deux horizons ne contenant presque point de diatomées et en ont encore un beaucoup plus riche. La Fig. 1 représente la coupe stratigraphique générale des dépôts de fond du nord-ouest de l'Océan Pacifique, basée sur l'analyse

Vol 6 1959

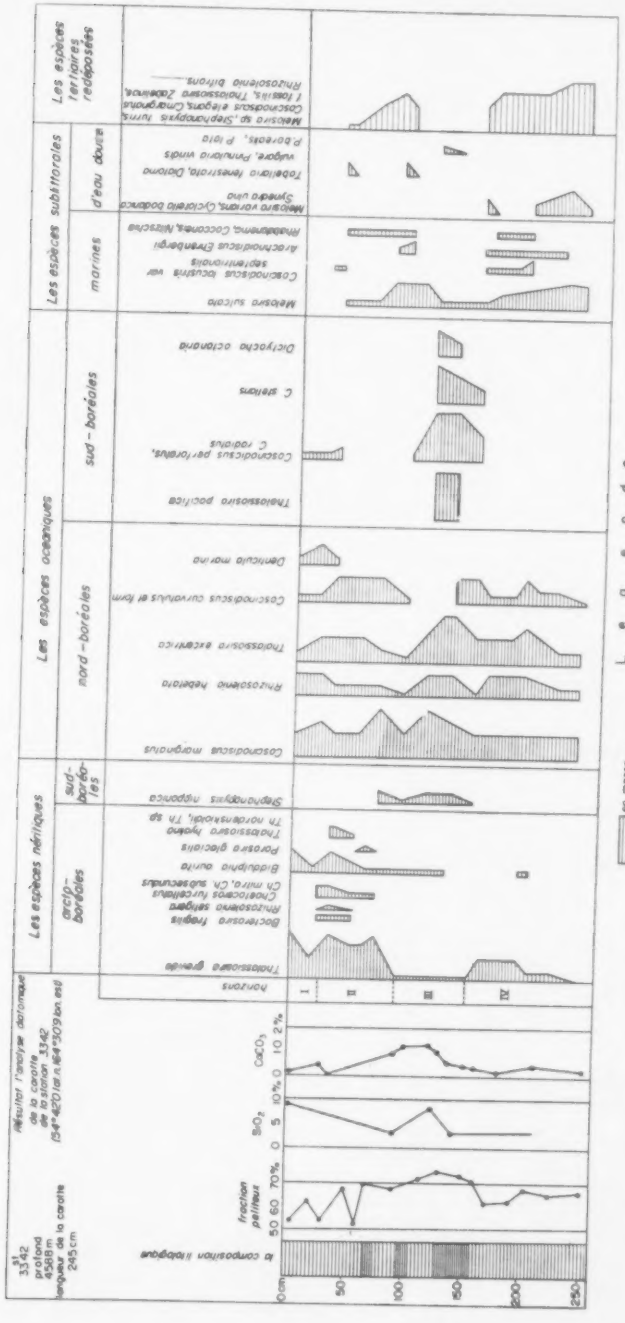

Fig. 2. Résultat de l'analyse des diatomées de la carotte de la station 3342.

des quatre carottes. La numération des horizons est la même que pour les mers de l'Extrême-Orient — en commençant par la surface elle est marquée par les chiffres I, II, III, IV (2. 3).

Dans les dépôts de l'horizon I (ou supérieur) dominent les espèces océaniques nord-boréales et arcto-boréales : *Thalassiosira excentrica* Cl., *Coscinodiscus marginatus* Ehr., *Rhizosolenia hebetata* Gran, *Denticula marina* Semina.

Les espèces néritiques ne prennent qu'une faible part dans la flore des diatomées et leur présence est signalée surtout aux stations situées près des bas-fonds continentaux et insulaires St. 3342 (Fig. 2) et St. 3378. A la station 3325, située sur l'élévation sous-marine Obroutcheff, on ne trouve point d'éspèces néritiques dans les dépôts de l'horizon I, dont l'épaisseur varie de 0·24 à 0·55 cm. Il est possible que l'épaisseur de l'horizon I aux stations 3325 et 3359 soit en réalité moindre, car la limite entre cet horizon et l'horizon inférieur reste indéterminée.

Dans les dépôts de l'horizon II prédominent (comme auparavant) les diatomées océaniques nord-boréales et arcto-boréales. Cependant on y trouve également les espèces arctiques sporadiques : *Melosira frigida* Jouse, *Rhizosolenia hebetata* f. *bidens* Karsten et *Biddulphia obtusa* Ralfs.

Il existe en outre certaines espèces néritiques qui entrent dans le plankton del a mer de Bering et de la mer d'Okhotsk (*Thalassiosira gravida* Cl., *Th. hyalina* Cl., *Th. kryophila* Cl., *Chaetoceros furcellatus* Cl., *Biddulphia aurita* Bréb. et Yodey).

En outre, dans les dépôts de la couche nommées ci-dessus on trouve de nombreuses diatomées, radiolaria et spicula d'éponges anciennes (des périodes géologiques reculées) dans le gisement redéposé. La limite inférieure de l'horizon II, contrairement à sa limite supérieure, est nettement marquée grâce au brusque changement de la flore.

Le maximum des diatomées se rapporte aux dépôts de l'horizon III. Nous y signalons une augmentation sensible de silice amorphe et de carbonate de calcium par rapport aux dépôts supérieurs et inférieurs. La flore des diatomées de l'horizon III, au nord-ouest du Pacifique, est caractérisée par les espèces sud-boréales. On trouve dans ce complexe : *Thalassiosira pacifica* Gran et Angst, *Coscinodiscus asteromphalus* var. *centralis* Grun., var. *subbuliens* A. Cl., *C. perforatus* Ehr., *C. radiatus* Ehr., *C. stellaris* Roper. De plus, la flore de cet horizon (surtout celle de sa partie supérieure) est caractérisée par l'espèce largement boréale — *Stephanopyxis nipponica* Gran et Jendo.

A côté des diatomées tempérées on y rencontre aussi des silicoflagellées (*Dictyocha octonaria* Ehr.) et des algues Péridinées.

Les diatomées nord-boréales communes, provenant du Pacifique (*Thalassiosira excentrica* Cl. et *Coscinodiscus marginatus* Ehr.), sont répandues dans les dépôts de l'horizon III, sous un aspect morphologique particulier. La première espèce y est représentée par des formes d'une dimension exceptionnelle, le diamètre de la valve étant de 65-75 μ. La deuxième espèce n'a point la striation de bord si typique des *Coscinodiscus marginatus* Ehr. contemporains. En outre, nous signalons des modifications dans la structure des aréoles de cette espèce. Les éléments tertiaires redéposés n'ont pas été trouvés dans les dépôts de l'horizon III. Les régimes de température les plus favorables au développement de la flore des diatomées mentionnées ci-dessus sont les eaux à température 12°-14°. Des températures analogues sont propres aux eaux de surface près de la limite sud de la région boréale (40°-42° de latitude nord).

Vol 6 1959

La composition de la flore de l'horizon III est semblable à celle de la mer du Japon, quoiqu'elle en diffère par quelques traits particuliers.

L'épaisseur des dépôts varie de 0·50 à 0·95 m.

Dans les dépôts de l'horizon IV les diatomées n'apparaissent que rarement. Les formes redéposées y prédominent, surtout les diatomées du Miocène. Nous y trouvons également une grande quantité de Radiolaria et de Spicula des éponges tertiaires. Le plus grand nombre d'éléments redéposés se trouvent dans les dépôts de la station 3342, près des Iles Commandors. Plus on s'éloigne de la côte moins on trouve de diatomées anciennes.

La flore synchrone aux dépôts est caractérisée par les diatomées océaniques et néritiques nord-boréales et arcto-boréales. Il existe également des formes sublittorales sporadiques marines ou d'eau douce. Par exemple à la station 3342 on a découvert 14 diatomées fluviales.

Des espèces arctiques — les mêmes qui ont été signalées pour l'horizon II — donnent à la flore de l'horizon IV un caractère froid (*Biddulphia obtusa* Ralfs, *Rhizosolenia hebetata* f. *bidens* Karsten). L'épaisseur visible des dépôts de l'horizon IV est de 0·97 m à 150 m.

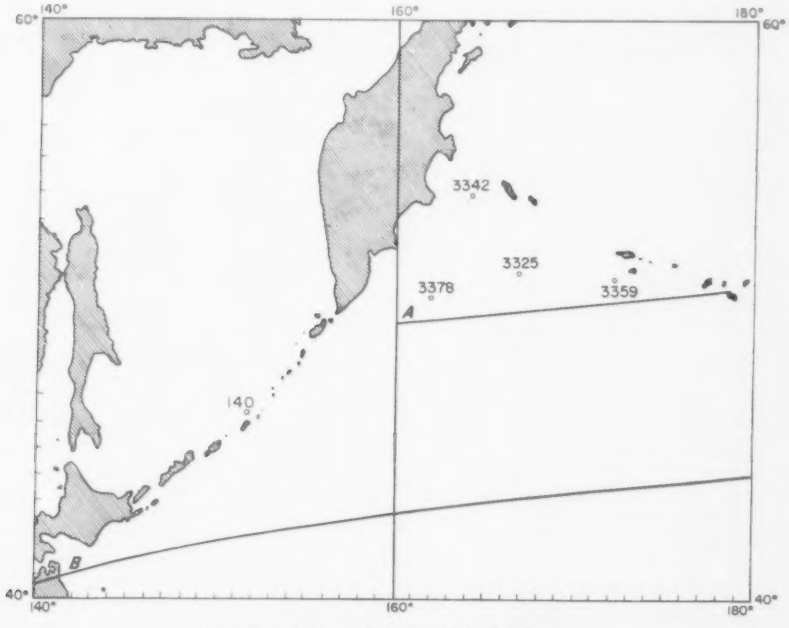

Fig. 3. Limite sud de la région boréale :
A Limite sud de la region boréale a l'époque interglaciaire
B Limite sud de la region boréale a l'époque actuelle

A la verticale des carottes l'on voit que le changement de la flore des diatomées se produit d'une manière synchronique et régulière. Des traits de ressemblance caractérisent les diatomées dans les dépôts des horizons I, II, III et IV de toutes les carottes. La flore de l'horizon I correspond à la flore actuelle du Pacifique, à l'est

du Kamtchatka. La flore des dépôts des horizons II et IV est plus froide. Une flore beaucoup plus tempérée que la flore actuelle caractérise les dépôts de l'horizon III. Sa composition est conforme au régime de température des eaux près de la limite sud de la région boréale, qui était déplacée évidemment vers le nord de 1000 km environ (Fig. 3).

Ainsi les diatomées dans la coupe verticale des carottes reflètent deux époques de refroidissement du climat et une époque où le climat est devenu plus doux.

Les régularités de la répartition des diatomées et la modification de leur composition floristique dans les dépôts de la partie nord-ouest de l'Océan Pacifique et dans la mer d'Okhotsk et de Bering sont semblables (2, 3).

Cela permet de synchroniser les dits dépôts et de reporter les horizons II et IV aux époques glaciaires et l'horizon III à l'époque interglaciaire.

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Vol 6 1959

A median valley of the Mid-Atlantic Ridge

M. N. HILL

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Abstract—A median valley of the Mid-Atlantic Ridge at least 75 miles in length running in a North-South direction is described. Its bottom lies at a depth between 1800 and 2200 fm and the flanks rise steeply to between 900 and 1200 fm. The width between the flanks is approximately 12 miles.

No direct evidence of the rock types in the area is available but the seismic evidence suggests the existence of basaltic lavas on the flanks. Evidence for present-day volcanic activity in the area comes from a measurement of the heat flow through the floor of the valley; this was about six times higher than is usual in deep sea or continental areas.

The valley was not associated with any large magnetic anomaly; it is however possible that its axis corresponds with the line of earthquake epicentres through the area.

Vol.
6
959-60

INTRODUCTION

DURING survey cruises of H.M.S. *Challenger* in 1953 (GASKELL, HILL and SWALLOW 1958), an area of the Mid-Atlantic Ridge between the latitudes 46° 30′N and 47° 45′N and the longitudes 23°W and 28°W was crossed in an East to West direction a number of times and continuous echo-sounding records were obtained on the Asdic echo sounder. The general features of this area are not well known, and no close surveying has previously been undertaken, and it was hoped that detail of at least part of the area could be obtained during the last expedition of H.M.S. *Challenger* in September 1953. A programme for obtaining detail over a comparatively small area was arranged and a start was made on it; before its completion, however, weather conditions became impossible for further progress. When the storm abated no time remained for further work.

In 1954 it was decided that R.R.S. *Discovery II* should return to the area for further detailed topographical work. It was hoped that as well as echo-sounding in the area it would be possible to obtain rock samples by coring or dredging. This proved impossible although short cores of the soft sediments existing in limited areas were obtained both by R.R.S. *Discovery II* and by H.M.S. *Challenger*.

A further expedition to the area was made in R.R.S. *Discovery II* in 1956 (HILL, 1957a) and measurements of the total force of the earth's magnetic field were made in the same area as in the 1954 expedition. Unfortunately further topographic work was prevented by adverse weather conditions during this 1956 expedition.

In order to obtain information about the structure of the area, seismic refraction shooting experiments were undertaken by H.M.S. *Challenger* in 1953 (GASKELL *et al.* 1958) and by R.R.S. *Discovery II* in 1956. A core of approximately 10ft. length and a measurement of the heatflow through the sea floor in the median valley were also obtained in the latter expedition. Further attempts at dredging rocks from the area failed.

(2) TOPOGRAPHY

The area of the investigations undertaken by H.M.S. *Challenger* in 1953 is shown in Fig. 1. This area is topographically unlike the Mid-Atlantic Ridge and Eastern Atlantic Basin further to the South. This area has been described by TOLSTOY (1951). He divides the Ridge into three regions as follows (a) a high central zone of the main range which is described as a series of parallel ridges following the general trend of the Mid-Atlantic Ridge. (b) an intermediate zone flanking the main range to the East and to the West between 1600 and 2500 fm which is particularly conspicuous in the region of the Azores. TOLSTOY names this region the Terraced Zone, and he notes that there are two types of transition between successive terraces. 'Some are separated by a step, with no intervening peak, whereas in others there is a peak or hump on the outer edge of the higher terrace.' (c) an outer mountainous zone between the abyssal plain and the terraced zone.

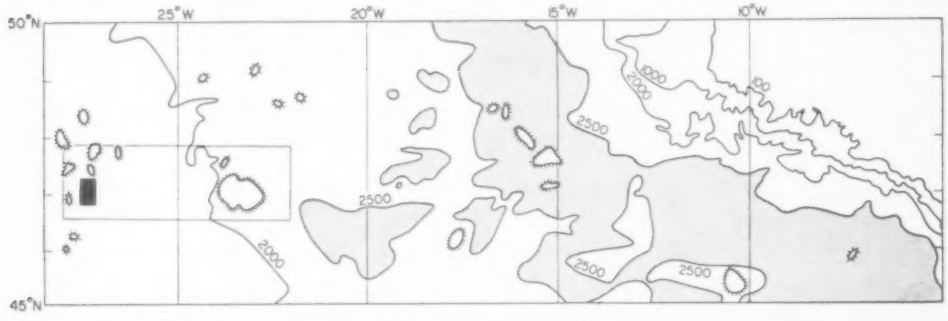

FIG. 1. Bathymetric chart of part of the Eastern Atlantic Basin. Soundings in fathoms. Contours are drawn at 100, 1000, 2000, 2500 fm. The area described lies within the rectangle drawn at the western side of this chart. The small black rectangle within the larger one is the area of detailed survey. The continental shelf off Brittany lies at the eastern side.

In the area shown in Fig. 1 this classification does not appear to be applicable. First, to the North of the Biscay Abyssal Plain further extensive abyssal plains do not seem to exist (HILL, 1956) and the area is not uniformly flat. There are undulating slopes and sharp pinnacles of rock but no extensive plains with the flatness of the typical abyssal plains further to the South. The area to the East of the Mid-Atlantic Ridge in these latitudes appears throughout to belong to TOLSTOY's 'Terraced Zone.'

The high central zone is a clearly defined feature but there is little evidence for this consisting to any great extent of parallel ridges. Fig. 2 shows the sections obtained in 1953 and the lack of correlation between them is conspicuous. The topography could not be accounted for by parallel ridges unless, between neighbouring sections, there are considerable alterations of direction. This seems improbable since the dimensions of the features in an East-West direction are not much less than the distances between neighbouring sections. It appears that there is a general slope up towards the high central zone and that this more violent topography is caused by isolated sea mounts which probably have a tendency for their North to South dimensions to be longer than those in an East to West direction.

Vol 6 1959

The conspicuous feature of Fig. 2 which appears on all the sections is the valley near the western ends. This valley is in an area where the peaks rise to minimum depth of about 700 fm although most of the peaks have depths around 1000 fm in a mean water depth of about 1200 to 1400 fm. The bottom of the valley is, throughout the length to which it has been followed, greater than 1800 fm and has a

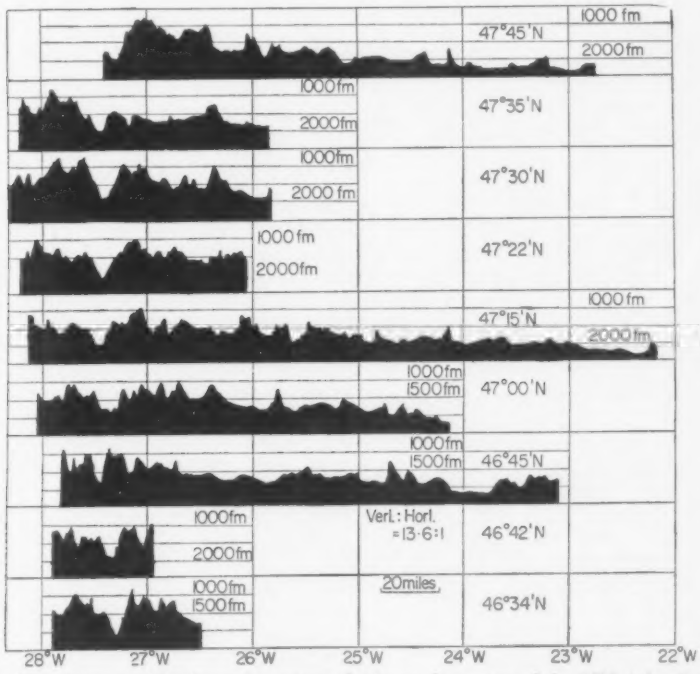

FIG. 2. East-West profiles from the eastern flanks to the centre of the Mid-Atlantic Ridge.

maximum depth of approximately 2200 fm; its width is variable but between the crests on either side it is of the order of 10 miles. The bottom of the valley which is comparatively flat is about 5 miles in width. Elsewhere in the high central zone of the Ridge in these latitudes there are no depths as great as those found at the bottom of the valley. The General Bathymetric Chart of the Oceans shows a few deep soundings which have probably been obtained from this valley but its conspicuous features have not, as far as we are aware, been previously reported.

The 1954 expedition was largely concerned with the survey of a small area within the larger area of the previous year's work and its position is shown in Fig. 1. The method of survey consisted of traverses along lines of latitude sixteen miles in length and separated from one another by two miles. Closer spacing might have led to difficulties on account of the uncertain North to South set of the ship which would become of progressively greater importance as the spacing between the East to West sections was reduced. Further, because of the beam angle of the sounder each section covers a strip of approximately $\frac{1}{2}$ mile width, the precise value depending upon the depth of water. Fig. 3 shows the sections obtained across the valley. The most northerly was on the run up from Britain and not exactly from East to West. There-

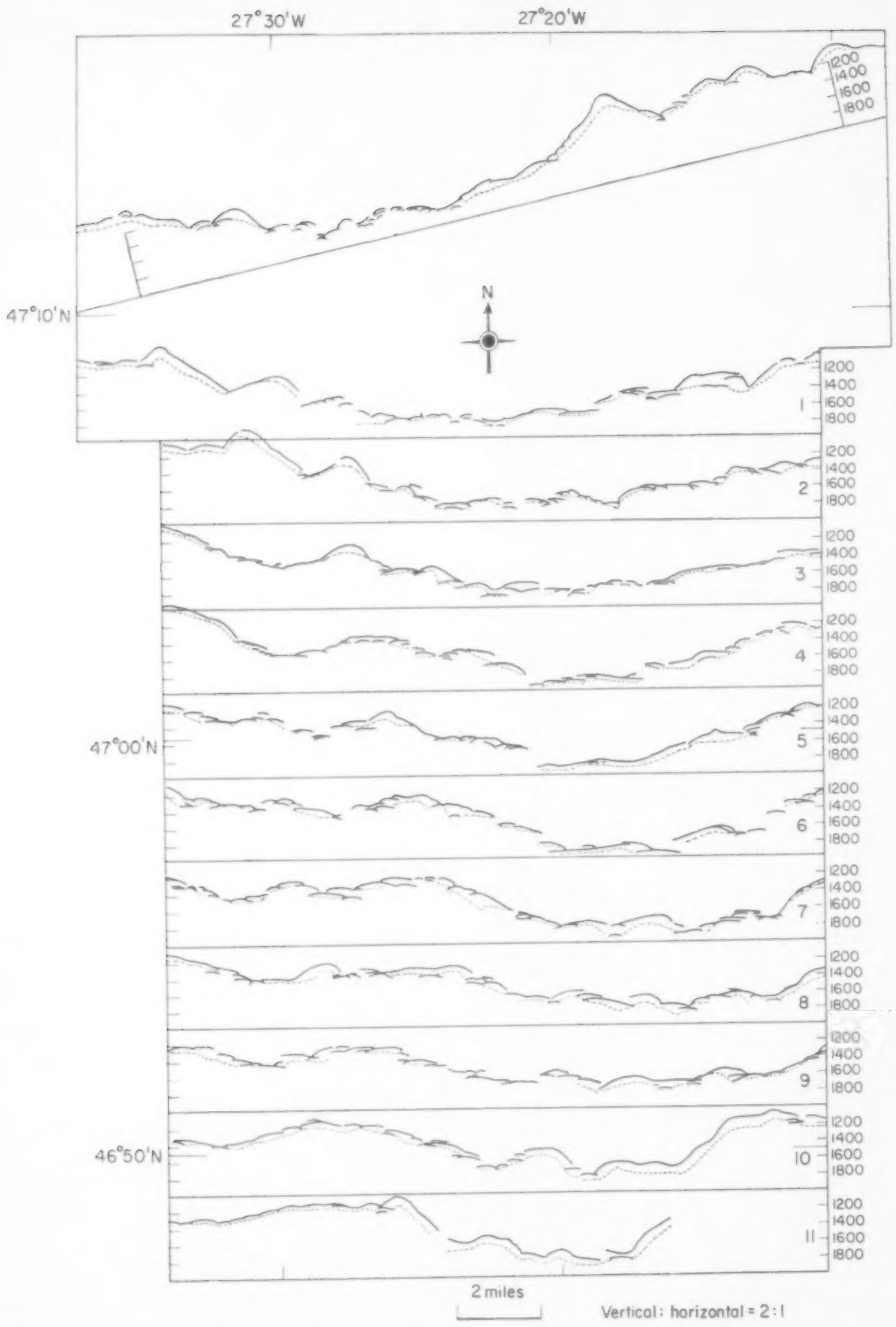

FIG. 3. East-West profiles through the deep valley. Soundings in fathoms. The profiles are separated from one another by 2 miles, except for the most northerly one which was at an inclined angle.

after the sections were along lines of latitude. The distance along each section was measured with the ship's log. It was impossible to obtain sights for the determination of the turning points. Sights obtained before and during the survey suggest that the uncertainty of position at any point in the area is less than two miles in any direction. Relatively the position of the sections will be in error by distances smaller than this. radio aids to navigation in this area are unsatisfactory.

On the sections shown in Fig. 3 the full details of the sounding records after correction for the variation of the speed of sound in water (MATTHEWS 1939) have been drawn. These show a considerable vertical spread with several clear-cut echoes from different apparent depths frequently being present at the same place. This complexity of sounding record results from the half power beam angle of the Kelvin Hughes Type 26E sounder used in this survey being approximately 8°. This angle allows, in areas of rugged topography, specular reflections from surfaces not vertically below the ship; in other words, any surface within the beam which is normal to the incident sound will provide a reflection. These multiple reflections result in uncertainty of the depth beneath the ship and consequent difficulties in selecting the appropriate depths for the purposes of contouring. We have in most cases selected the first reflection as representative of the depth although in a few places when the ship was over the bottom of the valley other reflections have been adopted since the first reflections had characteristics suggesting that they were coming from the sides of the valley. For example in Section No. 6 the depth of the eastern side of the bottom of the valley has been taken as the later reflection throughout the distance that this reflection appeared. Similarly in Section No. 3 the deeper reflection has been taken as representing the true depth at the western end of the bottom of the valley.

On these sections the dotted lines represent the time after which no further reflection returned. The time duration for which reflections occur is an indication of the roughness of the area within the sound beam although the apparent vertical extent of the reflection is greater than the true vertical scale of the roughness because of the beam angle of the sounder. This duration also depends upon the slope of the surface providing reflections; reflections from a steep rough slope will last for a longer time than those from a horizontal surface of equal roughness.

On the steep sides of the valley the value of the slope sometimes exceeds the beam angle. In this event specular reflection is not possible and the reflections are long drawn out, diffuse and without a sharp beginning; they are dependent upon the roughness for their existence.

The sections shown in Fig. 3 show various conspicuous features. Firstly there is the valley itself which reaches the same maximum depth within 100 fm on each crossing. Secondly, running parallel with the valley, there is a small central ridge which stands between 100 and 200 fm above the valley floor and which is between 1 and 2 miles across. Thirdly, there is the cliff-like western edge to the valley floor conspicuous in sections 4, 5 and 6. Lastly, there are the secondary depressions on the western slopes of the valley. Some of these features are more clearly shown in Fig. 4, the contoured chart of the area. In this chart it is apparent that although the area lying at depths greater than 1800 fm is continuous from North to South it becomes very narrow in the neighbourhood of sections 9 and 10. In sections 4, 5 and 6 it is considerably wider than elsewhere with depths exceeding 1900 fm; in these sections the mid-valley ridge exceeds 1800 fm in depth.

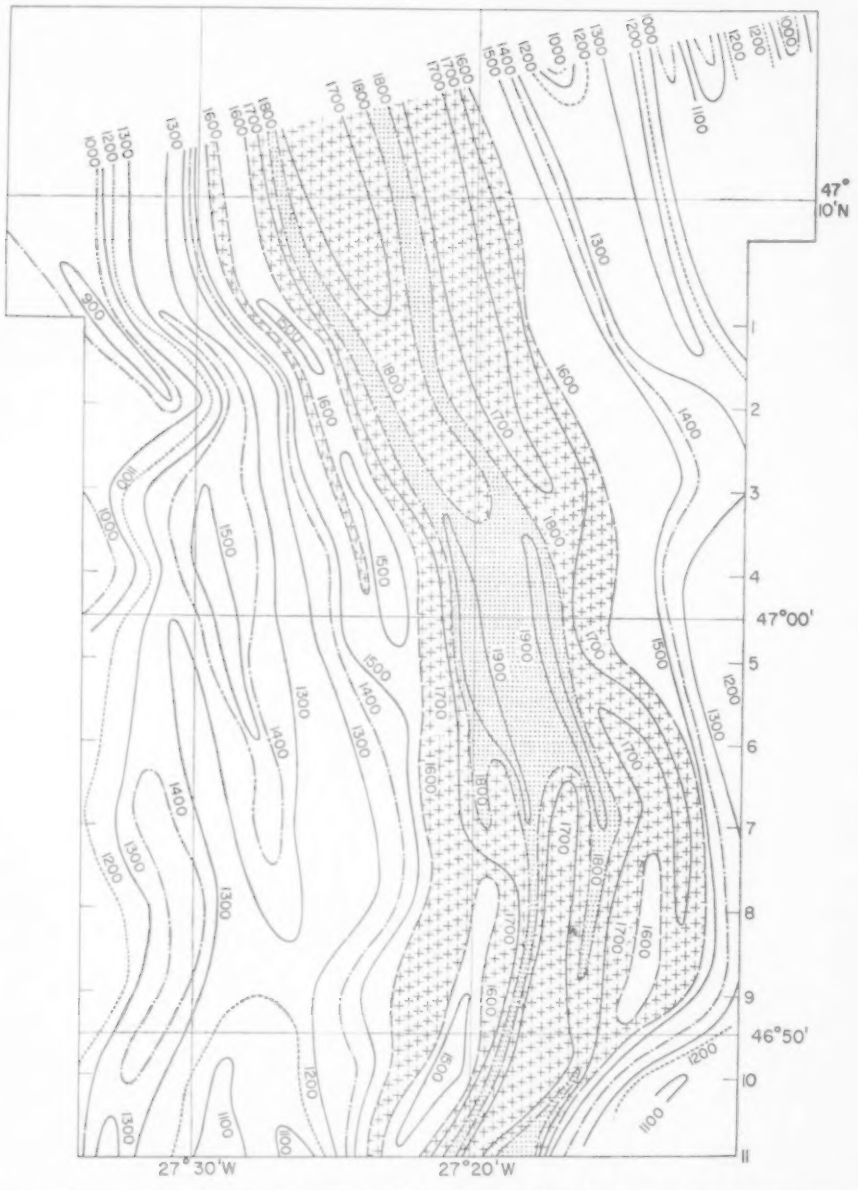

FIG. 4. Contour chart of part of the deep valley. Soundings in fathoms. The numbers down the side of the figure correspond with those of the profiles in Fig. 3. The shaded area has a depth greater than 1800 fm.

The uniformity of depth along much of this central channel suggests that it may be scoured by turbidity currents. The existence of the greater depths is evidence, however, against this hypothesis.

The two depressions on the western slopes are probably isolated from one another although it is possible that East-West channels interconnect them with one another and with the main valley.

In drawing the contoured chart of the area considerable assistance was derived from the short North-South runs connecting the sections. These showed that there were no breaks in the flanking ridges and from the character of the echo-sounder record, which was remarkably different from that obtained on the traverses, it was apparent that these ridges were comparatively smooth in a direction parallel with the valley itself. The valley was not, therefore, an elongated depression within a terrain of randomly spaced volcanoes.

Although the flanking ridges in the area of Fig. 4 were continuous this was not true further to the North. Thus in the unfinished detailed survey obtained in 1953 it was clear that there was an East-West break at least at one place on the western flanks. The contour chart obtained from the survey is shown in Fig. 5. The sections

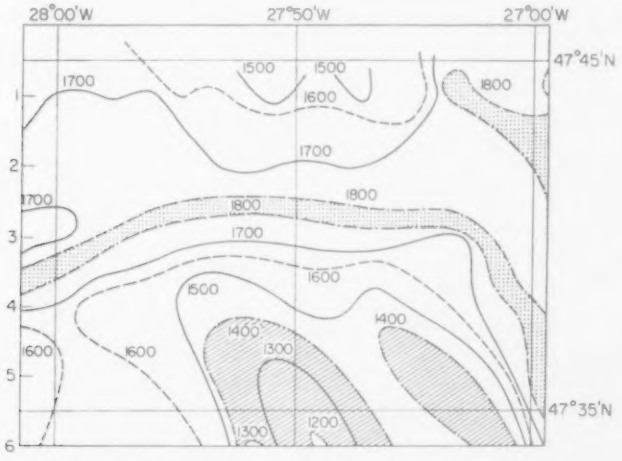

FIG. 5. Contour chart of an area to the West of the deep valley. Soundings in fathoms.

were East-West and 2 miles apart. The valley is at the eastern side of the area shown in this figure and the floor of the valley appears to be connected with a ' tributary ' running out to the westward. This feature was obtained from section 3 but it is not certain that the 1800 fm East-West contour goes right across the area as drawn. The depths at both ends of this section are, however, greater than 1800 fm and in both section 2 and 3 the depth always exceeded 1600 fm and generally 1700 fm. Even if the 1800 fm contours do not cross the area there is still no doubt of the existence of this break in the western flanks of the valley.

During the expedition of 1954 it was hoped that the valley could be followed to the South by steering a zig-zag course. This track is shown in Fig. 6; it started near the centre of section 11 of Fig. 4. In Fig. 6 the dotted area is that in which the depth exceeded 1600 fm and the dashed lines represent the 1200 fm contours.

On the second south-easterly run the valley was not crossed and it was therefore decided to steer due West. What is believed to be the valley was found further West than had been expected although the subsequent ship's track shown on Fig. 7 shows clearly that if this supposition is correct then the valley ends at latitude 45° 52′N.

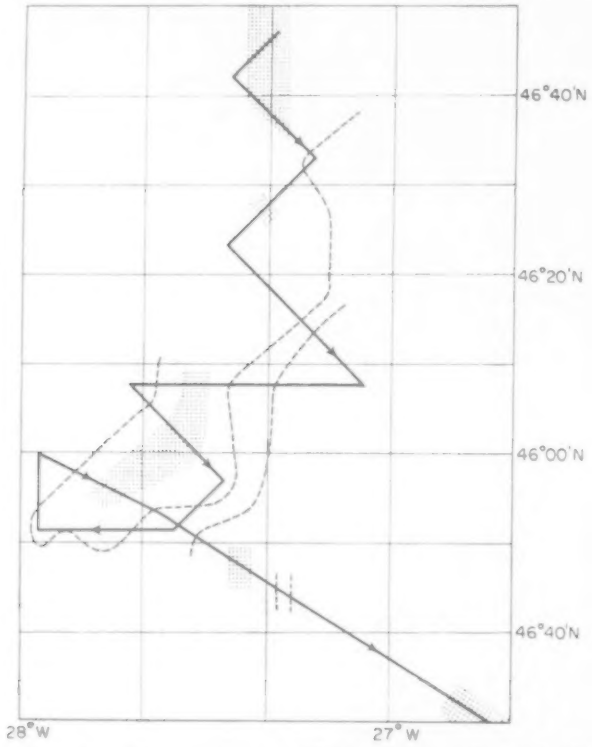

FIG. 6. The track of the ship after an attempt to follow the deep valley to the South. The dotted areas have depths greater than 1600 fm. The dashed lines are the 1200 fm contours.

Shortage of time prevented further investigation : it is however possible that the valley exists further South or that it is continuous but lies further West than the area searched. To established with certainty where the valley lies it would be necessary to obtain the depth profile along the 28°W longitude between 45° 40′N and 46° 20′N and an East-West profile along 46° 20′N from 28°W.

We have at present no indication of the northern limit of the valley.

(3) MAGNETIC MEASUREMENTS

During the 1956 expedition to the area we were unfortunate with the weather. Conditions were such that continuous echo soundings could not be obtained and navigation was difficult. Nevertheless the position of the deep valley was obtained on the down-wind runs to the eastward and interpolation was possible between these runs.

The weather did not interfere with the magnetic measurements of total field obtained from the nuclear-spin magnetometer designed and built in the Department

Vol 6 1959

of Geodesy and Geophysics at Cambridge (HILL, 1959). The total field contours of the area are shown in Fig. 7. On this chart the position of the valley bottom is indicated by the dashed lines and within these lines the valley depth is greater than 1600 fm. There is uncertainty of the position of this survey which might be as great

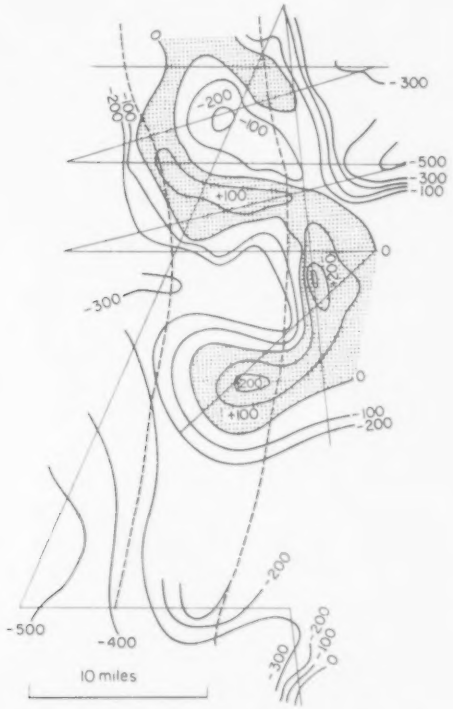

FIG. 7. The total magnetic field anomaly over part of the deep valley. The contour interval is 100 γ; the zero contour has a field value of 47,800 γ. Positive anomalies are shaded. The position of the deep valley is indicated by the dashed line and the ship's track by the thin continuous lines.

as 4 miles. For this reason no lines of latitude and longitude are shown. The centre of the area is however approximately 46° 50′N, 27° 20′W. This position corresponds to a point near the southern end of the survey shown in Fig. 4. The magnetic contours are shown after the subtraction of 47,800 γ from the field values. The ship's track is indicated by the thin black lines.

The results of the survey indicate that the deep valley is neither associated with any obvious magnetic anomaly, nor does there appear to be any aligment of the magnetic anomalies with the topography. Further, the magnitude of the magnetic anomalies within the neighbourhood of the valley are no greater than is found elsewhere in the Mid-Atlantic Ridge. This conclusion conflicts with evidence produced by HEEZEN (1958); he believes that this median valley, which he supposes to be more or less continuous along the Mid-Atlantic Ridge, is associated with a positive magnetic anomaly.

(4) SEISMIC REFRACTION SHOOTING MEASUREMENTS

On two occasions seismic refraction measurements have been made near the sides of the valley. The first occasion was in 1953 and the results from this work have been reported in GASKELL *et al.*, (1958). The second occasion was in 1956. With both experiments sono-radio buoys were anchored near the tops of flanking seamounts and shots were fired along lines stretching approximately North-South through the buoy positions. The technique of using sono-radio buoys has been previously described (HILL, 1950).

The following are details of the station positions, etc. :—

Station No.	*Position of Buoys*	*Depth under buoys*	*Length of line*	*Direction*
CR 42	46° 43′N, 27° 09′W	1000 fm	7 miles	330°
DY 28	46° 23′N, 27° 12′W	840 fm	8 miles	025°

With station CR 42 the superficial rocks did not provide refracted arrivals but it was concluded that the velocity of compressional waves could not exceed 3.8 km/s. Further, since the topography showed the area was made from solid rock the velocity was unlikely to be less than 2·5 km/s. Assuming the higher limit for the velocity the thickness of the layer (it was supposed that the layering followed the topography) would be 1·1 km; with the lower limit the thickness would be 0·6 km. Below this rock with unknown characteristics there was a layer with velocity approximately 5·7 km/s. No depths or velocities were obtained from any layer below this.

The results from station DY 28 showed that if there was any low velocity layer on the surface its thickness was less than 130 m and again because of the steepness of the topography it seems more plausible to assume that it does not exist. On this hypothesis the surface rocks of the area have compressional wave velocity of 3·6 km/s and are 1·2 km thick. The velocity below this is 6·4 km/s.

These results agree reasonably well with one another; it seems that the flanking mountains of the valley are covered with approximately 1 km of a low velocity rock which could well be a vesicular lava; below this there are rocks which could be basalts with velocity near 6 km/s. There is no evidence that these mountains are made from the so-called basaltic rocks which are usually found about 1 km below the deep ocean floor and which have a compressional wave velocity of 6·7 km/s. They might, however, be similar to the layer which is often found in the deep ocean between the unconsolidated sediments and the basaltic rocks. The velocity limits of this layer are usually taken as 4-6 km/s (HILL, 1957b); the two measurements we have obtained are near the high limit for this layer.

Confirmation of this interpretation is obtained from seismic experiments on a seamount North of Madeira where successful dredge hauls of basaltic rocks were obtained. The results from this area are unpublished. The two surface layers on the seamount had velocities of 3·7 and 6·2 km/s respectively. These values are similar to those found on the Mid-Atlantic Ridge although the thickness of the uppermost layer was 2·8 km.

Some dredging in the area of the deep valley produced a collection of stones which largely consisted of limestones and acidic igneous rocks; none of these could with certainty be attributed to the locality and may have been rafted. Similarly

Vol 6 1959

Vol.
6
959-60

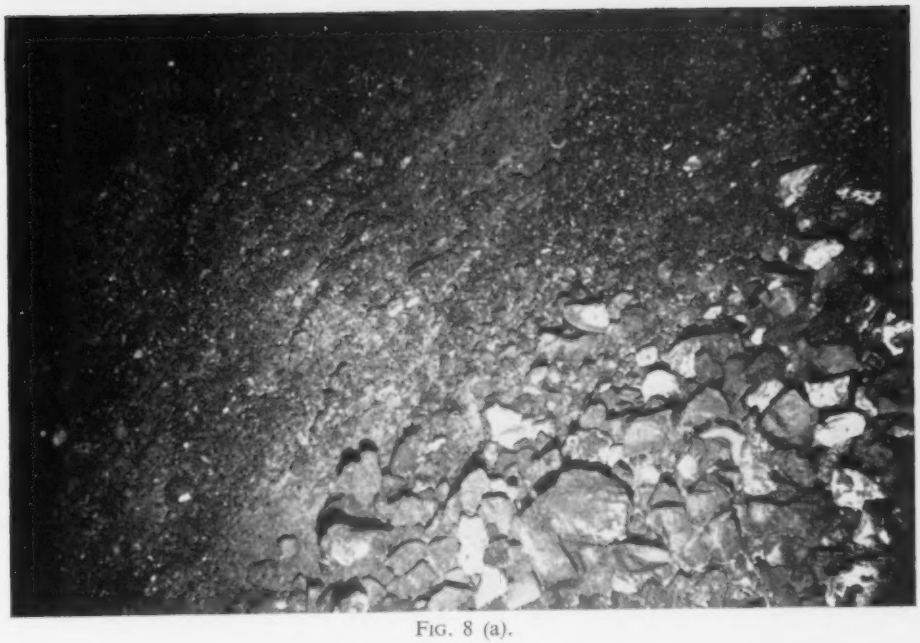

FIG. 8 (a).

FIG. 8 (b).

FIG. 8. Photographs taken near the top of the seamounts flanking the deep valley, showing the existence of angular cobbles.

Vol
6
1959

the coarse fraction of some cores taken in the valley bottom was analysed by Dr. A. A. DAY who concluded that it was representative of the volcanic and gneissic terrains of Britain, Iceland and Greenland and again might have been rafted. It is however strange that no rocks or particles of the basalt lavas which we believe to constitute the rugged topography of the area have been recovered.

Further efforts at dredging in the area are desirable and photographs taken near the top of one of the mountains flanking the valley show that they should be rewarding. Two of the photographs taken by Dr. A. S. LAUGHTON of the National Institute of Oceanography are reproduced in Fig. 8; these show angular cobbles which do not have the appearance of having been rafted and which therefore are probably of local origin.

The nearest direct information of the rocks of the Mid-Atlantic Ridge comes from areas hundreds of miles to the South where numerous basaltic boulders have been obtained (EWING, 1948). These have been described by SHAND (1949).

It is unfortunate that seismic measurements have not been obtained over the base of the deep valley; these have not been attempted on account of the difficulties of mooring the sono-radio buoys in the great depth of water.

(5) SEISMICITY AND VOLCANISM

The Mid-Atlantic Ridge is seismically active and most of the epicentres lie along the centre line of the Ridge which in our area coincides with the line of the valley. The precision with which the epicentres can be determined is such that it is impossible to see whether there is an exact coincidence between the valley and the earthquakes in the area; of the four epicentres in the latitude of our investigation, none are further than 20 miles from the valley and this distance is within the uncertainty of the epicentral determinations. It would seem plausible to suppose that faulting associated with the earthquakes may at least in part explain the formation of the valley.

There is no direct evidence to indicate recent volcanic activity in this part of the Mid-Atlantic Ridge; we have however indirect evidence from one measurement (unpublished) of heat flow made in the floor of the deep valley by Sir Edward Bullard during the 1956 expedition. The result from this showed that the heat flow was approximately six times higher than was usual in ordinary deep ocean areas or in continental areas; this suggested the existence of high temperatures in the rocks near the surface of the sea bed.

Several attempts at further measurements in the valley failed since the heat probe failed to penetrate, presumably because of the presence of rocks on the sea floor.

(6) CONCLUSIONS

The topographic features of the deep valley may be summed up as follows :—

(a) Its length is at least 75 miles. The northern limit is unknown, but to the South there is a possibility that it ends in about latitude 46° 40′N. Throughout the distance it has been traced, it runs approximately North to South and it lies between the longitudes 27° and 27° 30′W.

(b) Its maximum depth lies between 1800 and 2200 fm throughout the length; the bottom is not flat but shows minor irregularities. In the area where detailed investigations have been made there does however appear to be a

flat channel of variable width which might represent the track of turbidity currents. The presence of greater depths along the channel is evidence against this hypothesis. The width across the bottom is approximately 5 miles. It is of note that any other roughly straight line 75 miles in length in the Mid-Atlantic Ridge would show much greater relief than exists on the floor of the valley. It is therefore unlikely to represent a down-faulted block. Further, for the same reason, it appears that volcanoes have not erupted through the valley floor since its formation.

(c) The flanks rise steeply with slopes of approximately 1 in 2 to depths usually between 900 and 1200 fm. The flanks form ridges running parallel with the valley. The distance between these flanking ridges is about 10 miles.

(d) At least in one place there is a transverse valley coming from the West; in general however the North to South structures dominate the topography.

(e) Elsewhere in this region of the Mid-Atlantic Ridge North to South structures of comparable magnitude do not exist, and from the information available it appears that any ridges which exist elsewhere cannot be longer than a few miles.

The magnetic information can be summarized as follows :—

(a) There is no obvious magnetic anomaly directly associated with the valley.

(b) In the neighbourhood of the valley the size of the magnetic anomalies are not significantly different from those found elsewhere in the Mid-Atlantic Ridge.

The seismic refraction shooting experiments in the area indicate that the material of the sides of the valley are similar, at least in respect of their compressional wave velocities, to volcanic seamounts found elsewhere. It seems probable, in spite of the absence of samples, that basaltic lavas constitute the valley flanks and the high value of the heat flow through the floor of the valley suggests that volcanic activity still exists.

Finally we have the approximate coincidence of the axis of the valley with the line of epicentres of earthquakes. If there is correlation between the existence of the valley and the location of the epicentres then, as HEEZEN has postulated, we might expect to find perhaps with some blockages, the deep valley extending throughout the length of the Mid-Atlantic Ridge.

Continental rift structures with similar characteristics to the valley are hard to find; it does however appear that the graben of the Rhineland are in width not far different and they extend in length at least for distances of the same magnitude as the length over which we have followed the valley. These graben, however, represent down-faulted blocks and there is evidence, mentioned above, that the deep valley is not of this nature. The African and Red Sea rifts are more extensive laterally and are again down-faulted blocks. Nonetheless the tensions in the crust which produced these rifts on the continents might have produced the deep valley. This has been suggested by EWING who supposes that the postulated Median Valley along the Mid-Atlantic Ridge swings eastward around South Africa, continues along the ridges

Vol 6 1959

of the Indian Ocean into the Red Sea Rift. Similarly it is possible that the broad Median Zone of Iceland, although laterally of great extent, lies on the continuation of the deep valley to the North. Geologists suppose this region to be a graben. (HAWKES, 1941).

It has recently been shown by GIRDLER (1958) that in the Red Sea Rift there is a positive gravity anomaly. It would be interesting to obtain similar gravity profiles over the valley. Likewise it is desirable to shoot refraction seismic lines along the length of the valley to determine whether high velocity rocks are closer to the surface than elsewhere in the Ridge.

Acknowledgements—I am greatly indebted to Dr. J. C. SWALLOW of the National Institute of Oceanography who first discovered the existence of the valley in 1953, and who assisted in all later expeditions to the area. I am also most grateful to my colleagues and shipmates for their part in collecting the information. Lastly, it must be apparent that without the most valuable assistance of the Hydrographer of the Navy and of the Director of the National Institute of Oceanography in allowing us facilities in H.M.S. *Challenger* and in R.R.S. *Discovery II* this work would have been impossible.

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Vol.
6
959-60

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Observations of squid at the surface in the NE Atlantic

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(*Received* 15 *April* 1959)

Abstract—An account is given of visual surface observations on squid in the NE Atlantic. These show that *Ommastrephes pteropus* Steenstrup, is continuously distributed throughout the surface water of the very extensive region between the English Channel, The Azores and The Canary Islands.

IN a previous paper (BAKER, 1957) a brief account was given of the methods that have been used in the R.R.S. *Discovery II* for the study of oceanic squid; there was particular reference to the use of an underwater camera for this purpose. The methods described provide information about squid well below the surface and have revealed their presence at certain places and certain depths, but more work is necessary before any general conclusions are reached on the habits and distribution in deep water of even the commonest species.

Meanwhile, because so little is known at present about the distribution of squid in oceanic regions and because of their importance in the general balance of marine populations, it is felt that some facts about their observed distribution in a large oceanic area are worth putting on record. The information is supplied mainly by visual observations of the occurrence of squid at the surface at night, made in a series of ten cruises by the *Discovery II* and in two cruises of the Marine Biological Association's R.V. *Sarsia* in the north-east Atlantic between 1952 and 1958. These cruises, which were primarily for other oceanographical work, took place in spring, autumn or summer, and each lasted about 3–6 weeks.

The identification of squid seen at the surface cannot be definite ; those caught in handnets or on a rod and line have, however, been identified as *Ommastrephes pteropus** Steenstrup, 1855. Figs. 1–3 show that they have been caught at stations from all parts of the area in question and as they did not appear to be different from the rest of the population while in the water it is probable that most of the squid observed in the NE Atlantic in the way described below belong to the same species.

The stations at which surface observations of squid have been made by members of the staff of the National Institute of Oceanography are plotted according to seasons in Figs. 1–3. At stations at which no surface observations have been made but at which the presence of squid below the surface has been shown by the under-water camera or vertical long lines the position is marked by a triangle. Because a number of the observations were carried out on or near banks the 1000 fms (1829 m) contour is shown although it is not known whether depth influences the horizontal distribution of oceanic squid.

*There is considerable uncertainty regarding the use of the names *Ommastrephes* d'Orbigny, 1835, and *Sthenoteuthis* Verrill, 1880, and until this problem of nomenclature can be thoroughly investigated with additional material I follow the practice of recent authors in using *Ommastrephes* as the generic name for this common Atlantic form.

Vol
6
1959

Squid are usually seen only at night when the ship is stopped and the sea is illuminated by floodlights or cargo clusters. Those caught have varied in mantle length from a few inches to some 2 ft (60 cms), though full details of length frequencies of the specimens are not yet available. The squid may appear in the illuminated area either singly or in shoals. It was noticed that those tending to form shoals were mainly in the smaller size groups and their size did not appear to vary greatly within any one shoal. Because these groups move in and out of the illuminated area it is difficult to assess how many different shoals have been observed and also to

FIG. 1. Stations in the NE Atlantic at which observations have been made on squid. The depth contour is that for 1000 fms (1829 m).

assess the number of individuals in a shoal. In consequence only the presence or absence of squid is considered in this paper. The quantitative observation of squid at the surface is anyway of doubtful significance since the underwater camera and vertical long lines have indicated the presence of squid down to depths of 547 fms (1000 m) even when some were visible at the surface. It would be valuable if other methods of observation would allow the estimation of the numbers or biomass of squid/unit surface area.

Although, as the charts show, squid have been recorded as apparently absent from the surface at night on twenty-three occasions, the significance of such negative records depends to a great extent on the conditions at the time, particularly the state of the moon and the weather. This was apparent in 1956 when the *Discovery II*

was on station in approximately 40°N, 20°W from 13–30 September and from 27 September to 2 October. During the first of these periods surface observations were made on six consecutive nights and although a few small squid were seen and one was caught, the impression was obtained that the region supported only a small population. When the ship returned to the position on 27 September however, squid were found to be plentiful, forty-one being caught with hand-nets and rod and line in three nights. During the first week the moon did not set until well after midnight and, except for short intervals when there was thick cloud, the sea surface

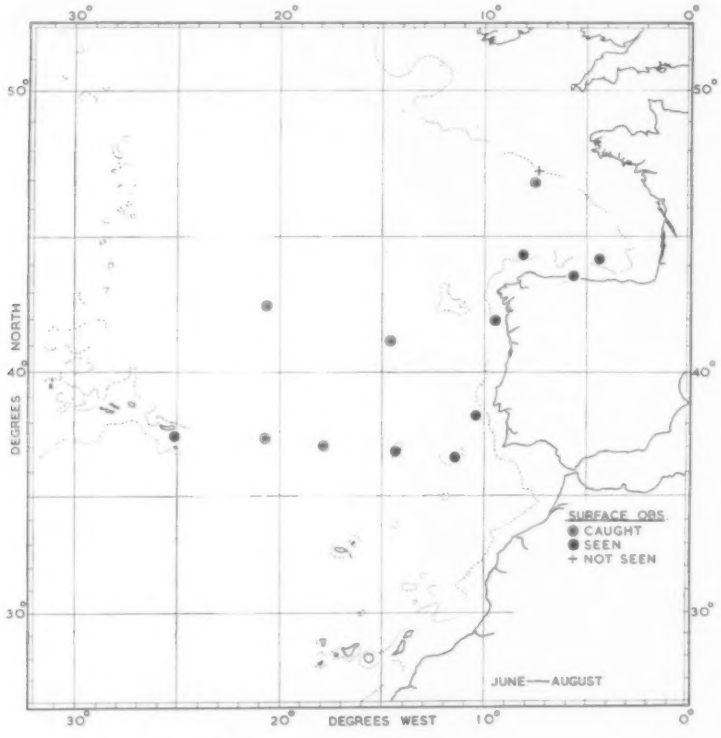

FIG. 2. See legend for Fig. 1.

was brightly illuminated, whereas by 27 September the moon was not rising until after midnight. It has been found that it is usually during the period between 2200 hr and midnight that most squid are seen and it is suggested that the apparent variation in the population noted at 40°N, 20°W was due largely to the presence or absence of moonlight during these hours.

It is obvious that the squid are attracted to the illuminated area, but it is not clear to what extent they may be attracted by the light itself or by small fish which are frequently also present. Small garfish (*Scombresox*) and myctophids were surprisingly abundant within a few inches of the surface, and the squid were often seen to devour both forms. Against the light shining above the water, these fish may be clearly visible to squid some way off. It may be that both light and prey

have the effect of drawing squid towards the ship from some little distance. In any case a bright moon would presumably tend to diminish the effect of the lights on both squid and fish.

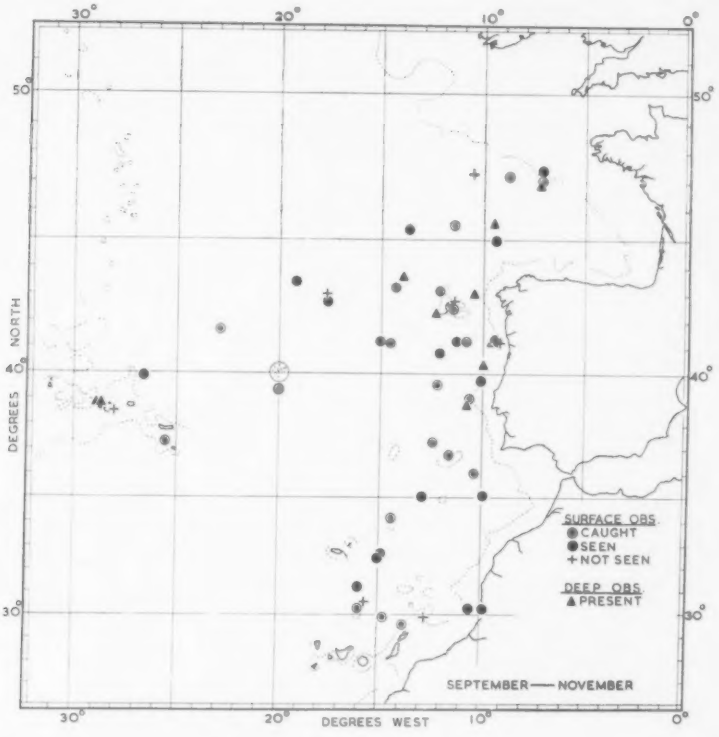

FIG. 3. See legend for Fig. 1.

The state of the sea surface is also an important factor affecting the reliability of records of absence, because observation becomes very difficult in rough weather.

Table 1. *The number of stations at which surface observations were made on squid in each season.*

	Spring (*Mar.–May*)	*Summer* (*June–Aug.*)	*Autumn* (*Sept.–Nov.*)	*Winter* (*Dec.–Feb.*)
Present	21	13	48	2
Absent	15	1	7	—
Total	36	14	55	2

Table 1 shows the results of the surface observations. Each night spent near position 40°N, 20°W has been counted as a separate observation, but for the sake of clarity they have been shown on the chart (Fig. 3) as one surface observation. Most of the observations were made in spring and autumn and only a few during the months of June, July and August. The two winter records were made while the ship was on

ocean weather station Kilo (45°N, 16°W) in February 1955 and I am indebted to Captain H. O. L'Estrange for this information. REES and MAUL (1956 p. 269) record *O. pteropus* as 'very abundant during the summer months of July, August and September' near Madeira, and ADAM (1952, p. 104) includes two records for early June, one at 01° 30′S, 01° 20′E where 17 specimens were taken and the other at 7° 10′N, 14°10′W where two specimens were taken.

The area in which the *Discovery II*'s observations were made and in which squid were found to be abundant is only a small part of the known range of *O. pteropus*. REES (1950 p. 40) lists the British records, and among these he includes a specimen taken in a trawl off St. Kilda (57° 49′N, 8° 34′W) which appears to be the northernmost published record for this species. The known southern limit of the range was greatly extended when ADAM (1952, p. 104) published accounts of specimens taken in the south-east Atlantic at positions 11° 45′S, 12° 40′E and 13° 25′S, 12° 10′E. Along the western Atlantic coast the species is found from Nova Scotia to the West Indies (JOHNSON, 1934). It has also been recorded from the Pacific by the Arcturus Expedition which took seven small specimens in a dredge at a position 60 miles south of the Cocos Islands. (4° 50′N, 87° 00′W) (ROBSON, 1948).

Thus the observations plotted in Figs. 1–3 add little or nothing to the known range of the species. What they do show is the frequency and continuity of its occurrence in a very substantial ocean area, amounting to about 700,000 square nautical miles. Squid were almost always seen within a few minutes wherever the ship happened to stop at night when there was no bright moon or rough sea. This implies that within the area, squid are usually somewhere within say a few hundred yards or less of any given point. The small numbers of negative stations do not necessarily signify the absence of squid.

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Vol
6
1959

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Quelques traits remarquables du relief sous-marin en Mer Mediterranée et en Mer Noire

(A) Profondeurs de 5000 mètres en Méditerranée

(B) Rides géantes par 1900 mètres de fond en Mer Noire

H. LACOMBE

(*Received* 23 *October* 1959)

Abstract—Presented here are a few remarkable features of the floor in the Mediterranean and Black Seas, from soundings obtained in July and August 1955, on board the *Calypso*, under command of Cdr. J. Y. COUSTEAU. These soundings show the existence of depths of 5000 metres in the Ionian Sea, south of continental Greece, and of ripples reaching 40 mètres in height in regular depths of 1900 metres in the southern Black Sea, off the Bosporus.

Vol. 6 959-60

AU cours de la campagne hydrologique effectuée sous notre direction (LACOMBE et TCHERNIA, 1959) à bord de la *Calypso* en juillet et août 1955, divers accidents remarquables du relief sous-marin ont été reconnus, en particulier en Mer Ionienne, au large du Cap Matapan, dans la région des plus grandes profondeurs connues de la Méditerranée, et aussi en Mer Noire aux abords Nord Est du Bosphore.

(A) MER IONIENNE AU LARGE DU CAP MATAPAN

Le Commandant J. Y. COUSTEAU et le Docteur H. E. EDGERTON, qui se trouvaient alors à bord et disposaient d'une caméra pour grands fonds, désiraient prendre des photographies dans les profondeurs les plus importantes de la Méditerranée. La recherche de l'emplacement favorable à l'exécution des photographies fut dirigée par le Commandant COUSTEAU grâce à un réseau de sondes qui est reproduit sur la Fig. 1 en traits fins; le cartouche de la Fig. 2 donne la situation générale de la zone étudiée à l'WNW de la Crête et au SW du Cap Matapan.

A partir d'une série de points au radar sur le Cap Matapan, une estime soignée fut tenue pendant cette recherche, effectuée de nuit (25-26 juillet 1955). Son exploitation immédiate conduisit à choisir le point A (Fig. 1) pour prendre les photographies par 35° 38′ 4″N et 21° 48′ 7″E, près du raccordement d'une pente faisant face au SE et d'une zone de relief régulier. Une bouée fut mouillée en ce point sur câble de nylon et servit de repère pendant les délicates manoeuvres nécessaires à la prise des vues du fond. A l'issue des opérations, le lendemain, 27 juillet 1955, le navire, faisant route sur l'île de Cérigo, se trouva pendant quelques minutes au-dessus de fonds de 5,000 m.

Pour ce travail occasionnel, qui était une annexe à la mission principale d'hydrologie, la position du bâtiment ne fut pas l'objet de déterminations astronomiques systématiques et ne saurait être considérée comme exacte à mieux que 2 milles près. Aussi les informations qui sont données ici ne le sont-elles qu'à titre de contribution à la connaissance de la forme du fond et ne peuvent elles être considérées comme assez

précises pour être portées sur des cartes de navigation à échelle supérieure au millionième. La sonde maximum de ces parages (4600 m toutes corrections faites), pratiquement identique en grandeur à celle trouvée par le *Berlin** en 1928 (4632 m), est située seulement à 1·3 milles au Sud. La carte grecque No. 65 publiée en février 1952 donne aussi dans cette région des indications de fonds très comparables dénotant une dépression orientée WNW–ESE. L'accord de ces diverses sources est donc relativement étroit.

FIG. 1.

Les sondes ont été obtenues au moyen d'un sondeur EDO modifié, dont la grande gamme s'étend de 0 à 3000 brasses, et non de 0 à 6000 comme sur l'appareil de série; sur la planche I, la largeur de bande représente donc 3000 brasses et non 6000 comme l'indiquent les valeurs imprimées. Par construction, l'appareil est règlé pour une alimentation électrique en 60 périodes par seconde et une vitesse du son de 4800 pieds par seconde. La fréquence de l'alimentation a été fréquemment notée et sa valeur moyenne entre deux instants a été, pour le dépouillement, déduite de l'écartement des repères d'heure portés au cours de la sonde sur la bande d'enregistrement. La corection afférente à la région géographique a été faite ensuite conformément au document H.D. 282 de l'AMIRAUTÉ BRITANNIQUE (LACOMBE and TCHERNIA, 1959).

On voit (Fig. 1) que la zone sondée affecte la forme d'une vallée creusée d'environ 400 m dans le relief moyen et qui débouche dans un évasement dont la partie Nord

*La position et les circonstances du sondage du *Berlin* nous ont été aimablement communiquées par le Bureau Hydrographique International, à qui nous exprimons notre gratitude.

Vol 6 1959

Vol.
6
959-60

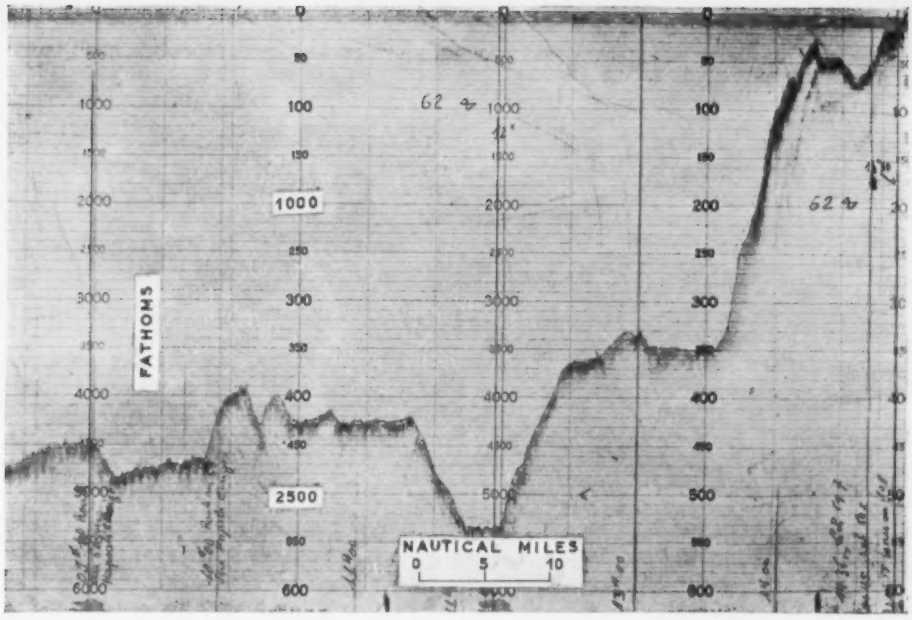

PLATE 1.

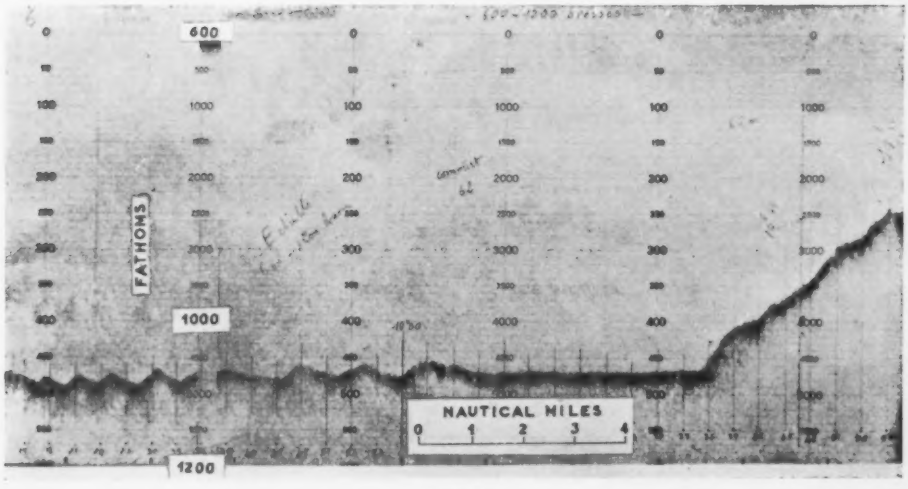

PLATE 2.

Vol
6
1959

seulement a été explorée. Le fond y présente un relief régulier où les profondeurs de plus de 4500 m occupent une zone elliptique s'étendant sur 5 milles selon le méridien et la moitié environ selon le parallèle. La vallée sondée qui ne présente pas de ramification, dans sa partie haute, est orientée WNW–ESE; elle est longue d'environ 8 milles et son thalweg, très sensiblement horizontal sur plusieurs milles, est recouvert de 4450 m d'eau. La largeur de la vallée est de l'ordre de 2 milles. C'est au Nord de la vallée et de l'évasement que la pente moyenne du fond est la plus importante et atteint environ 22°. Au SW de la vallée, le relief sous-marin semble franchement moins accidenté qu'au NE où se rencontrent des fonds supérieurs à 4000 m, à 7 à 8 milles au NE de la vallée, des fonds de moins de 3600 m étant interposés.

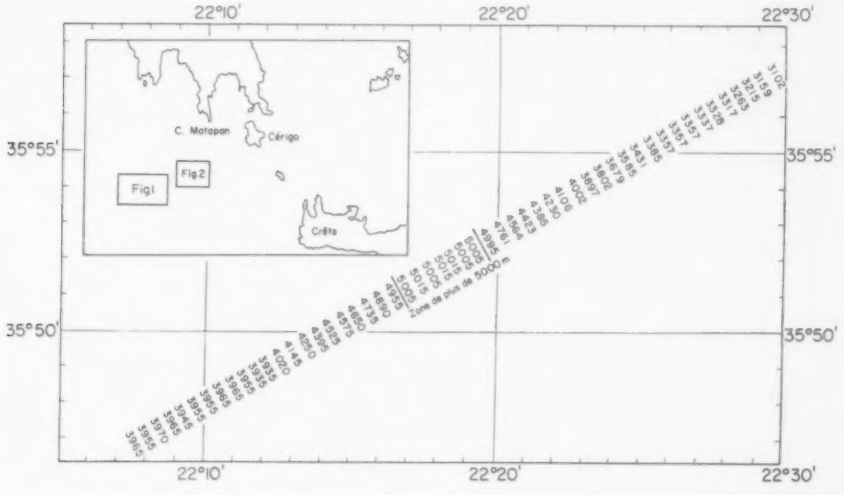

FIG. 2.

A partir du point A en direction 60° vers l'île de Cérigo (Cythère) la *Calypso* coupa (Fig. 2) une dépression très profonde située à environ 35 milles et 244° du Cap Trakili, pointe Sud de Cérigo. La sonde, toutes corrections faites, y est de 5005 à 5015 m; le centre en est situé par 35° 51′ 9″N et 22° 18′ 0″E, avec une précision de l'ordre de un mille. Le fond de la dépression (planche I) a 2·8 milles de largeur et est en moyenne horizontal; ses flancs remarquablement rectilignes présentent des pentes de 10° au NE et de 8° au SW qui la raccordent au profil général du fond, assez accidenté, par des dénivellations de 1500 m environ au NE et 1000 m environ au SW. La largeur de la dépression au sommet des pentes est de l'ordre de 12 milles.

C'est là, à notre connaissance, la plus grande profondeur jamais obtenue en Méditerranée. La plus grande profondeur classiquement connue était celle sondée par le *Berlin* qui est placée sur la Fig. 1. La carte greque No. 65 porte une sonde de 4850 m par 35° 44′N et 21° 29′E; ce serait donc une autre dépression car les fonds trouvés sur la ligne de sondes située au NW de la Fig. 1 ne mettent en évidence par plus de 4000 m aucune communication de cette dépression avec la vallée que nous avons sondée.

(B) MER NOIRE AUX ABORDS NORD-EST DU BOSPHORE

Les 7 et 8 août 1955 trois stations hydrologiques furent exécutées (LACOMBE et TCHERNIA, 1959) en mer Noire. L'une d'elles réoccupait sensiblement la station classique du *Thor* faite en août 1910 par 41° 32′N et 29° 24′E (point F, Fig. 3). Au cours des déplacements effectués dans cette mer, la *Calypso* fit des sondages systématiques dont les résultats essentiels sont représentés Fig. 3. Le segment le plus occidental EF était fait à la sortie du Bosphore en route au NNE, le second AB, au début duquel le contact radar sur la terre était encore acquis, fut effectué à l'issue des travaux variés (dont des prises de photographies du fond) exécutés à la station commencée en F. Une nouvelle station fut occupée en B; la fermeture sur la terre eut lieu aux environs de C sur Kili Bournou.

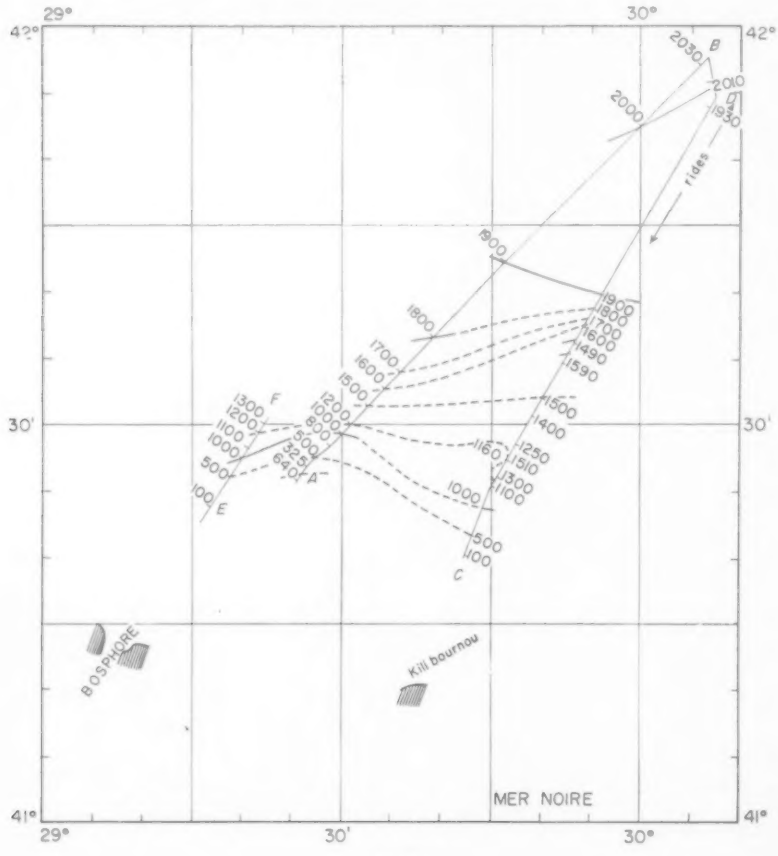

FIG. 3.

La Fig. 3 montre que le raccordement du talus aux fonds réguliers de 1900 à 2000 mètres de profondeur est franchement orienté NE–SW, c'est-à-dire en direction générale du Bosphore. Sur le trajet EF la pente du fond ne change pratiquement pas de sens, exception faite de quelques petites dépressions d'une profondeur propre faible

(15 à 20 m) présentes sur le plateau continental (dont le rebord est situé dans tout le secteur à environ 110 métres de profondeur) ou sur le talus; au contraire, on note sur AB une dépression au voisinage de A (640 m) alors que, au Nord, existent des profondeurs plus faibles (325 m); et sur CD, se rencontrent plusieurs dépressions dont les thalwegs sont situés à des profondeurs franchement plus grandes : 1300 par 41° 26′N, 1510 par 41° 27′ 5″N, entre 1160 et 1250, et 1590 par 41° 35′N. environ. En outre, comme le montrent les sections AB et CD présentées par la Fig. 4, il y a une importante différence dans la forme des profils rencontrés sur ces deux lignes; selon AB (Fig. 4), la pente diminue régulièrement du niveau A (325 m) au niveau B (2030 m);

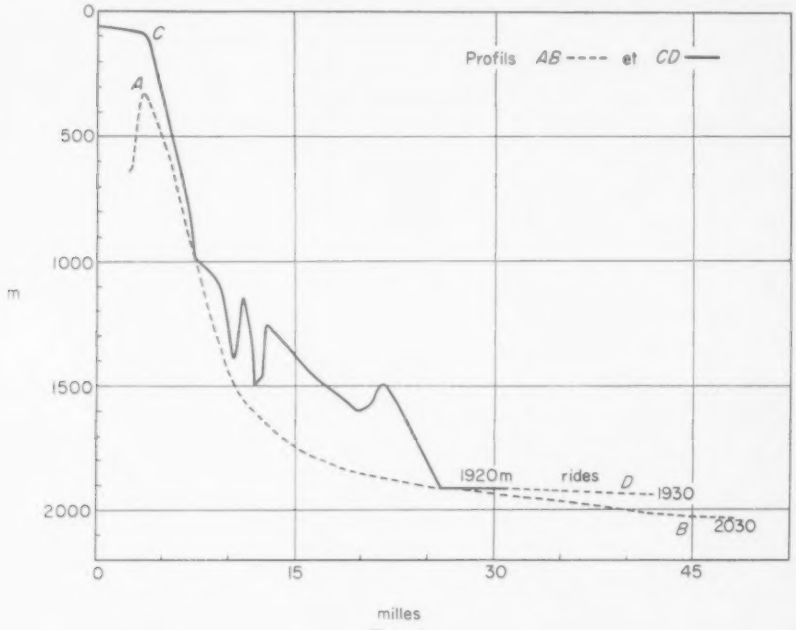

FIG. 4.

selon CD, on note d'abord la présence de trois vallées remarquables non pas tant par leur creux propre, qui est du même ordre que celui de la vallée au Sud de A, mais par le fait que les trois thalwegs se trouvent à des niveaux franchement plus bas que sur AB : ce trait suggère des creusements divergents à partir des abords de E, au droit du Bosphore. On remarquera aussi la discontinuité de pente au pied du talus selon CD, qui contraste nettement avec la continuité notée sur AB. La planche II montre la bande de sondage au voisinage de ce raccordement entre la pente d'environ 6° et une zone rigoureusement horizontale à 1920 m de profondeur. Une autre différence remarquable de la morphologie de la zone profonde selon AB et CD est l'existence de neuf " rides " sous-marines présentes sur une distance d'environ 9 milles au SW du point extrême du sondage (D) et séparées du pied du talus par la partie horizontale à 1920 m de profondeur, large d'environ 4·5 milles (planche II); il y a très probablement d'autres rides au Nord Est de D. Aucun phénomène de cet ordre n'apparait sur le profil AB, pourtant distant seulement de 5 à 8 milles selon les points. Ces rides ont des amplitudes notables (8 à 22 brasses, soit environ 15 à 40m)

et leur " longueur d'onde " évaluée sur le trajet CD oscille de 0·6 à 1·3 milles soit environ 1100 à 2400 m. Le versant situé au NE a toujours une pente supérieure à celle du versant SW, ce qui, sur ces rides, suggèrerait la présence d'un écoulement fluide orienté vers le NE (provenant du Bosphore dans certaines conditions au moins (?)). La variation de la longueur d'onde selon CD ne peut guère être attribuable à une obliquité plus ou moins grande du profil CD par rapport aux crêtes des rides qui, dans cette hypothèse, ne seraient donc pas rectilignes. En effet la longueur d'onde apparente passe pratiquement du simple au double par une transition brusque à hauteur de la 5 éme ride à partir de la gauche de la planche II (NE); il y aurait donc non pas des rides à crêtes courbes mais plutôt deux systèmes de rides, pratiquement parallèles, s'imbriquant l'un dans l'autre au NE de la 5 ème ride qui a d'ailleurs l'amplitude maximum (22 brasses, soit 40 m).

Acknowledgement—Nous tenons ici à exprimer notre gratitude au Commandant COUSTEAU qui, après avoir dirigé l'exploration des fonds au Sud du Cap Matapan, voulut bien mettre à notre disposition les bandes d'enregistrement et le journal de navigation de la *Calypso* et à M. P. GUIBOUT qui se chargea de la rédaction des sondages à notre Laboratoire.

Muséum National d'Histoire Naturelle, 43-45 *rue Cuvier*, *Paris* 5°.

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Vol
6
1959

On the abyssal circulation of the world ocean – II. An idealized model of the circulation pattern and amplitude in oceanic basins

HENRY STOMMEL and A. B. ARONS

(*Received* 15 *August* 1959)

Abstract—Stationary planetary flow patterns driven by source sink distributions in an ocean on the rotating earth were developed in Part I. These theoretical results are now used to construct a highly idealized model of the general abyssal circulation of the world ocean. The model is based on the postulation of the existence of two concentrated sources of abyssal waters (one in the North Atlantic and another in the Weddell Sea) and on a uniformly distributed upward flux of water from the abyssal to the upper layers as part of the mechanism of the main oceanic thermocline. Order of magnitude calculations based on this model lead to a variety of estimates of the time in which the deep water is replaced (from every 200 to 1800 years). Comparison of leading terms in the dynamical equations and equation describing the flux divergence of a tracer shows that there is a large range of lateral eddy coefficient which will influence the distribution of the tracer but not affect the dynamically determined planetary flow patterns.

Vol.
6
959-60

INTRODUCTION

IN Part I of this series of papers (STOMMEL and ARONS, 1959) we outlined the manner in which stationary flow patterns in oceanic basins on a spherical earth might be predicted for various source-sink distributions. These predictions were based on a previously published theory and interpretation of rotating basin experiments (STOMMEL *et al.*, 1958). In this paper we use the preceding results and a theory of the oceanic thermocline (ROBINSON and STOMMEL, 1959) to develop a model of the general abyssal circulation of the world ocean. The model, is of course, highly idealized, but we feel that it affords a plausible starting point for visualizing certain broad, qualitative features of the circulation, and that it may help generate suggestions for experiments and observations that will provide tests of the validity of the concepts which have been used.

OUTLINE OF THE MODEL

WÜST (1938) first presented a chart of the potential temperature at the bottom for the whole ocean, and showed that this clearly indicated that there was a powerful source of low temperature water in the Weddell Sea, and that this water spreads northwards along the western side of each ocean basin. This chart has become a standard exhibit in oceanographic textbooks, and, taken by itself, would suggest that on the basis of temperature alone all the deepest water in the ocean originates mainly in the Weddell Sea.

Of course, WÜST did not mean to suggest that there was only one source, and in a later paper (1951) he explicitly showed the great influence of the waters of the North Atlantic – through their high salinity – on the rest of the deep ocean water. This is also shown by the study of the T-S relations of deep water (STOMMEL, 1958) and by a study of the charts presented in Figs. 1(a) and 1(b), which we have compiled

partly from WÜST's similar chart of the Atlantic, and partly from a new analysis of data in the Indian and Pacific Oceans. (The spacing of isohalines in Fig. 1(a) is not uniform, but is laid out in geometric progression to prevent extreme crowding of lines in the North Atlantic). One can see that there is a very high salinity tongue extending from the Straits of Gibraltar, down the Atlantic and then eastward around Antarctica, and finally extending up northward into the Indian and Pacific Oceans – the salinity falling all the way. This chart serves to emphasize the importance of the contribution of North Atlantic Deep Water to the general abyssal circulation. The Mediterranean outflow is not massive enough to warrant inclusion as a third mass source. Its high salinity, however, may be regarded as supplying a 'tracer' by means of which the flow of water from the greater North Atlantic mass source can be followed.

Another figure which helps to visualize the sources of the two abyssal water components and the way in which they sink down and flow around Antarctica and thence up into the Indian and Pacific Oceans is presented in Figs. 2(a) and 2(b)*. Thus in Fig. 2(a) several vertical profiles of salinity are sketched to show the spread and vertical extent of the water at the salinity maximum already depicted in Figs. 1(a) and 1(b). To emphasize the features under description only two isohalines are shown: 34·70‰ and 34·76‰, although of course the total range of salinity variation in the ocean is very much greater than this narrow range. In the North Atlantic profile in Fig. 2(a), the entire profile is filled with water of salinity greater than 34·76‰. The North Atlantic is the source of the high salinity Deep Water. In the South Atlantic the high salinity at the surface is separated from that in the Deep Water by a layer of fresh Antarctic Intermediate Water.

We can trace the high salinity Deep Water all around Antarctica – but we note it does not extend all the way to the bottom at the very foot of the Antarctic Continent. A somewhat diluted portion (between 34·70 and 34·76‰) extends up into the Indian and Pacific Oceans at depth. The high salinities shown at the surface in the Pacific Ocean do not connect anywhere with the deep maximum. In the Indian Ocean the picture is somewhat more complex. There is a flow of saline water from the Red Sea which influences the salinity in the Deep Water of Equatorial and North Indian Ocean, and masks the influence of the North Atlantic component on the salinity there.

In Fig. 2(b), two isotherms are plotted on the same profile: 0°C and 2°C. The very cold water, less than 0 °C, originates at the surface in the Weddell Sea, and extends eastward around Antarctica, vanishing about at the Ross Sea, and never extending very far north. A more dilute portion, between 0°C and 2°C reaches further northward: in the Atlantic it occupies only a small portion of the deepest part of the western trough of the South Atlantic, but does not extend into the North Atlantic. This cold water occupies all the deep regions of the Pacific and Indian Oceans. In the North Indian Ocean it is clear that the Red Sea Water does not extend into the Deep Water on account of the temperature – although the salinity profile in Fig. 2(a) is ambiguous on this point.

In making the above description of the distribution of properties we have purposely used words like 'extends' instead of 'flows,' and 'occupies' instead of 'fills up,' in order to avoid implying that a process of actual flow is occurring. Actually, of course, the inference we wish to draw from the distributions of salinity and temperature in

*Previously unpublished, but constructed by E. D. STROUP and H. STOMMEL in the summer of 1958.

Vol 6 1959

Vol.
6
959-60

FIG. 1(a). Chart of the depth of the deep salinity maximum in the World Ocean, in metres. Aitoff equal-area projection.

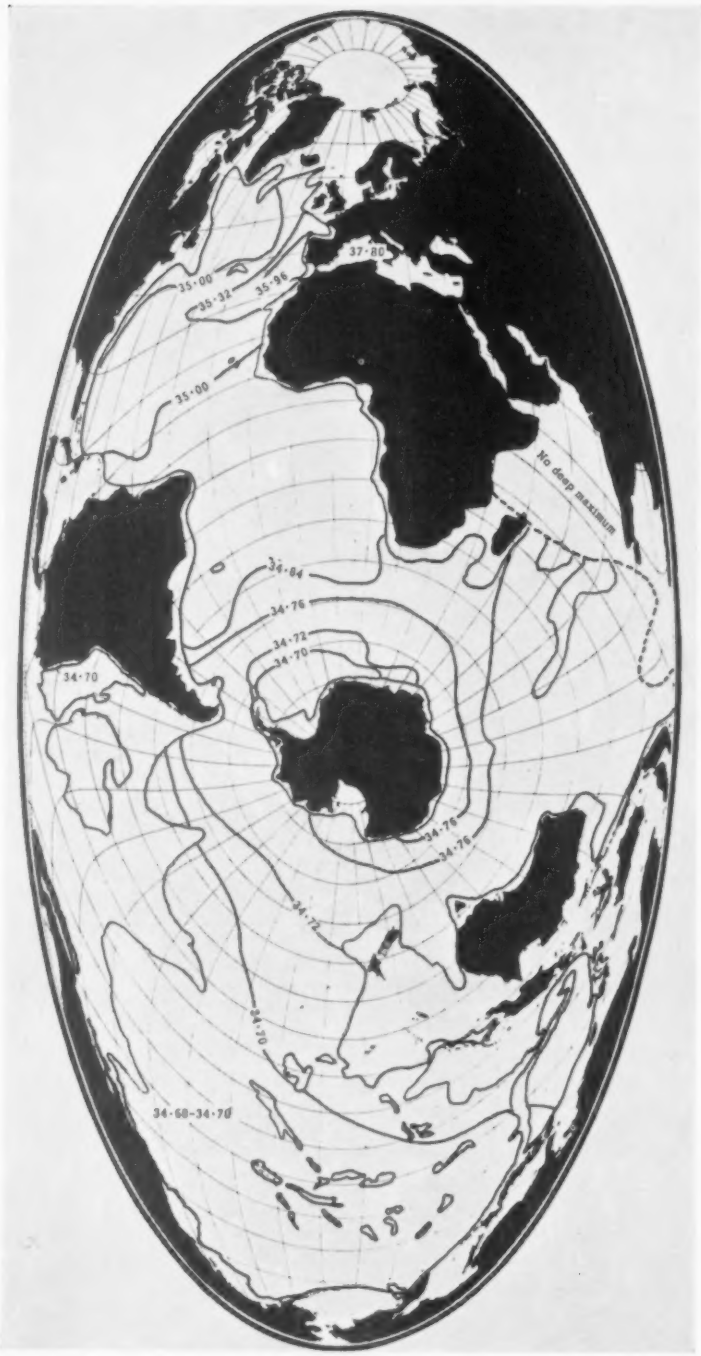

FIG. 1(b). Chart of the salinity, in parts per mille, at the depth of the deep salinity maximum. The isohalines are not spaced at equal intervals; high values near the Mediterranean are spaced more widely.

Vol
6
1959

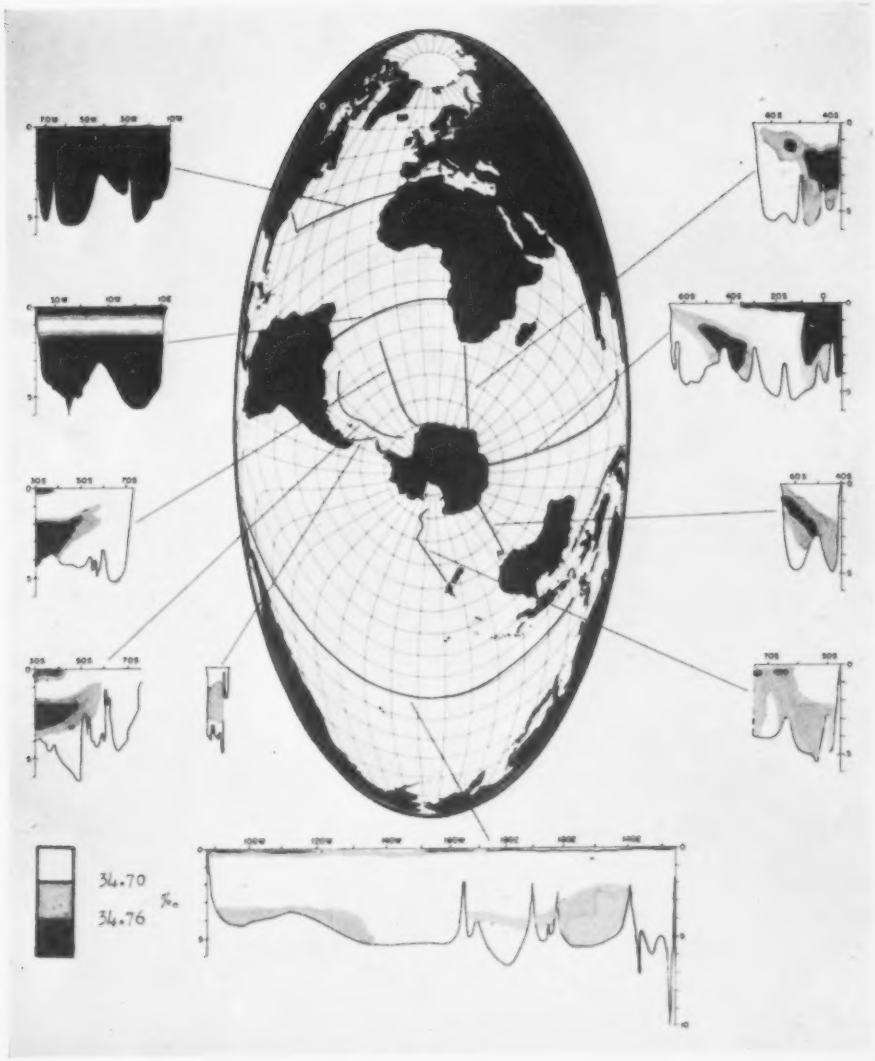

FIG. 2(a). Sketch of distribution of salinity on vertical profiles using only two selected isohalines chosen to depict in particular the influence of North Atlantic Deep Water on the Deep and Bottom Water characteristics of the rest of the World Ocean.

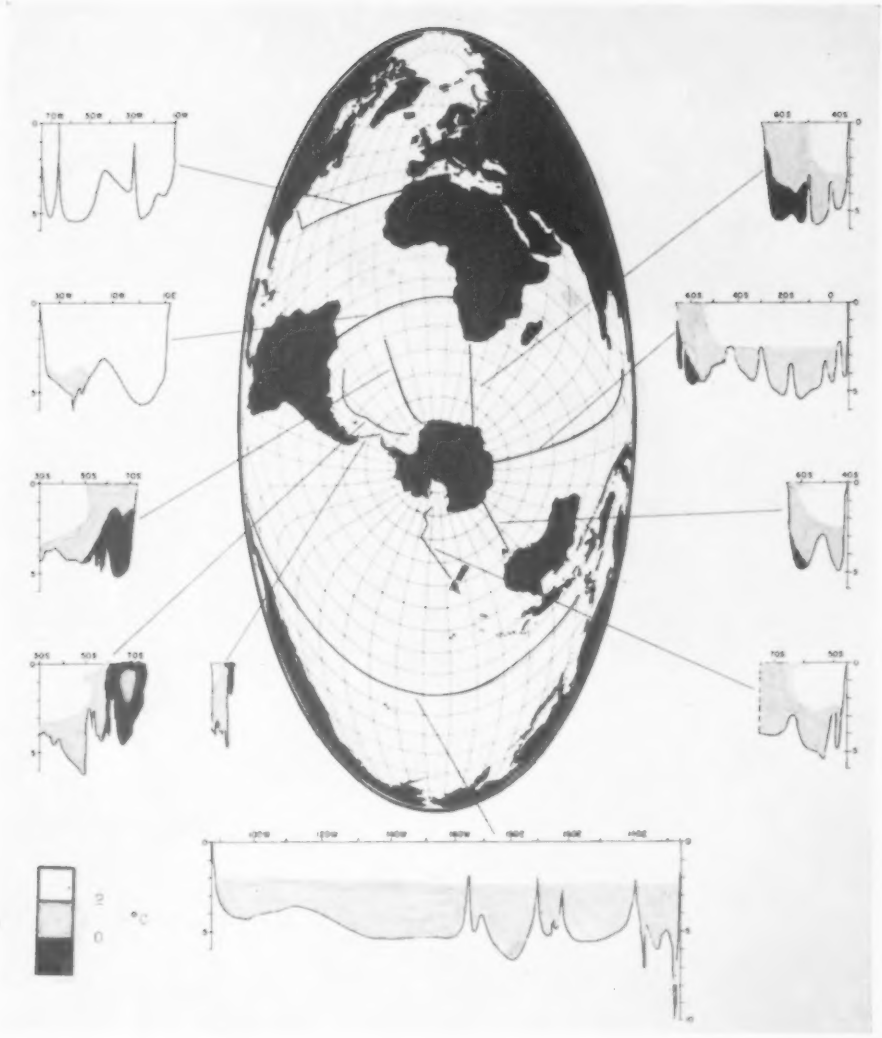

FIG. 2(b). Sketch of distribution of temperature on vertical profiles using only two isotherms chosen to depict the influence of cold Bottom Water from the Weddell Sea on the temperature of the Deep and Bottom Waters of the World Ocean.

Vol
6
1959

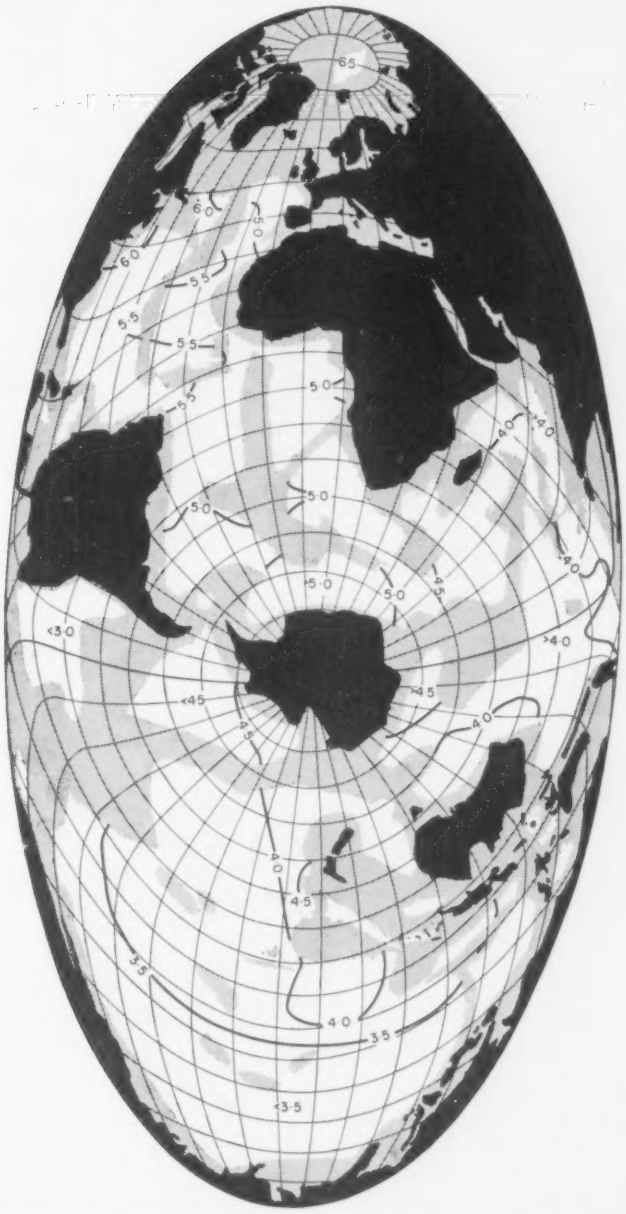

FIG. 3. Sketch of the distribution of dissolved oxygen (ml/l) in the World Ocean at a depth of 4000 m., showing elevated oxygen values in the western North Atlantic and near the Weddell Sea. The oxygen is depleted in regions far removed from these two source areas.

the deep waters is that they are evidence of flow, and this is the inference oceanographers have traditionally made on the basis of the tongue-like distribution of properties. But, as we will see, the tongue-like distribution of properties indicates flow only in a very general way and not in detail. Thus we believe that the distribution of properties merely helps us to indicate the location of the sources of the deep water and general direction of flow from one ocean basin to another. However, we believe that the detailed structure of the mean currents in individual ocean basins – the strong western boundary currents, and large cyclonic gyres – is not shown clearly by the distribution of temperature and salinity because of a masking effect due to lateral mixing – a subject into which we will enquire further in Section 5 of this paper, and will take up *in extenso* in Part III.

A chart of the concentration of dissolved oxygen at 4000 m is presented in Fig. 3. This chart is a preliminary edition of a set of deep ocean charts now in preparation by E. D. STROUP and STOMMEL, and is based upon the totality of deep oxygen measurements available. It shows the high oxygen content of the North Atlantic Deep Water, and the secondary maximum of oxygen in the neighbourhood of the Weddell Sea. Throughout all the rest of the ocean, the deep water oxygen decreases. Quite evidently there is no source of deep water in the Indian or Pacific Oceans.

Thus, in the light of the oxygen data, we assume that bottom water is generated by cooling in only two locations: (a) in the North Atlantic, (b) in the Weddell Sea. If in addition we assume that the mechanism of the oceanic thermocline involves a general upward motion of water at mid-depths, the theory presented in Part I leads to a system of boundary currents and interior geostrophic flow of the qualitative pattern sketched in Fig. 4.

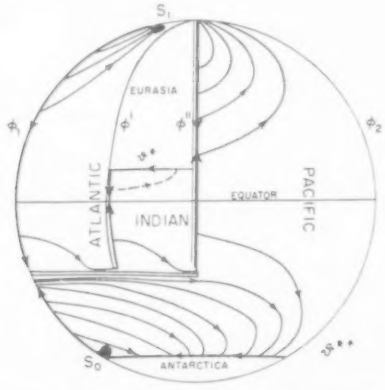

FIG. 4. A circulation diagram showing theoretical distribution of abyssal currents proceeding from the two sources S_0 and S_1 and flowing into a diffuse sink uniformly distributed over the whole ocean area.

Fig. 4 shows a sketch of a global ocean with meridional boundaries at ϕ_1, ϕ_2, ϕ' and ϕ'' and latitudinal boundaries at ϑ^*, ϑ^{**}. Thus we have very roughly a picture which looks like the real ocean basins. On the left is the Atlantic. The barrier ϕ' extending across the equator to mid-latitudes is the European-African land mass. The barrier ϕ'' is the East-Asian coast, extending down across the equator to Australia and New Zealand (in fact, of course this is not completely impermeable to water

Vol 6 1959

flow. The land mass bounded by ϕ', ϕ'' and ϑ^* is the Asian Continent. South of it we find the Indian Ocean. From ϕ'' to ϕ_2 we have the Pacific Ocean. The barriers ϕ_1 and ϕ_2 are the eastern and western edges of the North and South American Continental land mass drawn straight across Drake Passage to Antarctica. South of latitude ϑ^{**} lies Antarctica. Source S_0 is the source of Antarctic Bottom Water; Source S_1 is the source of North Atlantic Deep Water.

Isobars and schematic western boundary currents are sketched according to the principles described in Part I of this series – assuming a general upward motion over the entire ocean at mid-depths. From the North Atlantic source S_1 the water flows southward in a western boundary current. Part of this current feeds an interior geostrophic flow in the North Atlantic to the east and north-east. Some of this water rises vertically through the thermocline and the rest eventually rejoins the boundary current in higher latitudes nearer to the cource, thus forming a North Atlantic abyssal gyre.

It is well known that the existence of the southward flowing boundary current beneath the Gulf Stream was verified in the March–April 1957 *Discovery II – Atlantis* Expedition. Most of the North Atlantic Deep Water crosses the equator in the main strength of the western boundary current down the coast of South America past Brazil and Argentina to about 35 °S latitude. Along this course more water is lost to an interior geostrophic flow to the east and south-east, but we hypothesize that this water does not form a closed South Atlantic abyssal gyre, because upon reaching the latitude 35 °S it joins a zonal current flowing eastward toward the Cape of Good Hope. (See Section 7 and Fig. 16 of Part I). The zone bounded by 35°S and 55°S is marked by currents that flow all around the world; it is the zone in which communication between the Atlantic, Indian and Pacific Oceans is constrained to occur by the geographical distribution of the continents. We hypothesize that the Antarctic Bottom Water from the source S_0 flows northward along the western edge of the Weddell Sea to the 35 ° – 55 °S zone, where it veers eastward, joining the backing N Atlantic component, to form a great zonal stream that carries Atlantic water into both the Indian and Pacific Oceans. Upon passing the Cape of Good Hope some of this current flows northward in a narrow western boundary current that feeds the abyssal circulation of the entire Indian Ocean. The remainder of the zonal stream flows eastward, and after passing south of New Zealand, flows northward into the Pacific Ocean as a narrow western boundary current along the Tonga–Kermadec Trench. After crossing the equator the western boundary current flows northward to a latitude of about $30° \pm 10°$N where it exhausts itself. Analogy with the dynamical model shown in Fig. 8 of Part I suggests that further north in the N. Pacific there may actually be a southward directed western boundary current. South of latitude 55°S the pattern of flow seems to depend rather critically on what transport is permitted to flow through Drake Passage between the southern tip of South America and the northern tip of the Palmer Peninsula on the Antarctic Continent. It has been pointed out, however, in an earlier article (STOMMEL, 1957) that Drake Passage does not afford a free zonal passage for deep flows because of the position of the island arc on the eastern side. All except surface layers must (and indeed do) move northward in passing through. We do not have an adequate understanding of just what happens dynamically in this region, and as yet cannot introduce this complication theoretically in a quantitative manner. Therefore, in later sections the flow through

Drake Passage will have to be carried as an undetermined parameter. In Fig. 4 the rather drastic assumption that it vanishes has been made.

ESTIMATE OF THE AMPLITUDE OF THE ABYSSAL CIRCULATION ON THE BASIS OF TRADITIONAL OCEANOGRAPHIC INFORMATION

We have constructed in Fig. 4 a schema of the abyssal circulation, showing the patterns which would ensue if there is a general upward motion over the entire ocean at mid depths, fed by two equal point sources in the North and South Atlantic. We have not assigned an amplitude to this flow pattern. How much water is actually flowing in this pattern, in millions of cubic metres per second ? We now proceed to attempt to answer this question, first by comparison with the quantities of flow observed directly in the March – April 1957 *Discovery II – Atlantis* Expedition to the western boundary current in the North Atlantic, and secondly by a study and simplification of the dynamic calculations of the data recently published by WÜST (1955 and 1957). In the next section we shall determine the amplitude of the abyssal circulation by means of the new theory of the oceanic thermocline of ROBINSON and STOMMEL (1959).

We first consider the results of the joint British-American expedition to survey the deep currents under the Gulf Stream off Charleston, North Carolina, U.S.A., in the Spring of 1957. To date, only a short notice of the results has been published (SWALLOW and WORTHINGTON, 1957) and it is possible that estimates of deep flow will be rendered somewhat differently in the forthcoming more deliberate account of the survey which these authors are preparing, but if the depth of no axial motion in the western North Atlantic is about 1500 m., geostrophic calculations made on sections from Nova-Scotia to Bermuda and from Chesapeake Bay to Bermuda indicate that there are about 20×10^6 m^3/sec flowing southward in a narrow intense western boundary current along the coast of the United States. (However, when dynamic calculations of transport in the undercurrent off Charleston were made by WORTHINGTON and SWALLOW – as yet unpublished – much lower values were obtained : 3 to 6×10^6 m^3/sec). The transports computed by WÜST (1957) from the South Atlantic *Meteor* data show that the North Atlantic Deep Water flows southward along the Brazilian coast – riding over an intruding wedge of Antarctic Bottom Water – with about the same transport. Of course there is an arbitrary element in all dynamic current calculations based on hydrographic data alone – namely the choice of reference level – but the water mass characteristics in the South Atlantic are so very clear-cut that we are inclined to place confidence in WÜST's choice of reference level. We identify this narrow current of North Atlantic water with the western boundary current in our theoretical schema (Fig. 4) and think of it as merely being a somewhat attenuated continuation of the one under the Gulf Stream. Unfortunately we do not have reliable bases for estimating the deep flows in other oceanic areas, and so we cannot press this method further. From the T-S relations of deep water for the whole world it looks as though the contribution of flow from the Weddell Sea must be at least as much as that from the North Atlantic – at least another 20×10^6 m^{3}sec^{-1} so that the total rate of addition of water from the two sources S_0 and S_1 in Fig. 4 must be reckoned as at least 40×10^6 m^3/sec.

Having made reasonable guesses of the values for the two sources, we can now attach numerical values to all other parts of the circulation, and in so doing we

Vol 6 1959

naturally must make a crude allowance for the various dimensions of different oceanic basins. In Fig. 5 we have divided the world ocean into 14 portions :

North Atlantic into two sub-areas each of	2	unit areas
South Atlantic	2	2
North Indian	2	1
South Indian	2	3
North Pacific	2	5
South Pacific	2	5
Southern Ocean south of Africa	1	2
Southern Ocean south of Australia	1	2

The total area of the world ocean is thus divided into 40 unit areas. Inasmuch as the actual area of the 3000 m. surface in the world ocean is about 3×10^8 km^2, we must associate $7{\cdot}5 \times 10^6$ km^2 with each of the above unit areas. Although the

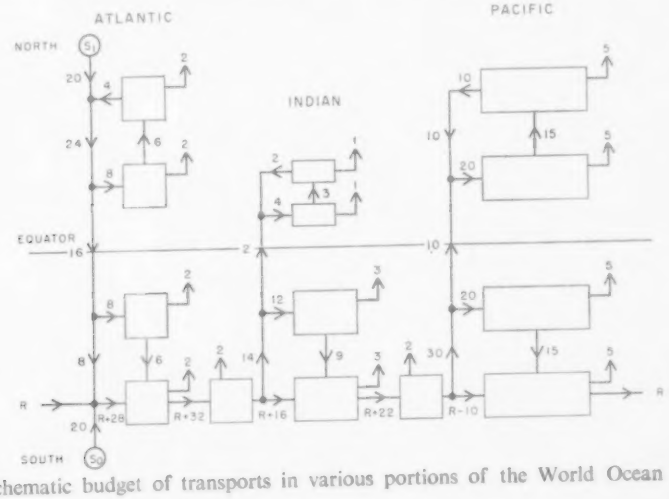

FIG. 5. Schematic budget of transports in various portions of the World Ocean (see text).

theory of the oceanic thermocline (ROBINSON and STOMMEL, 1959) indicates that the vertical velocities at mid-depths are functions of position we will, for simplicity, assume that the vertical velocity is nevertheless uniform over the ocean, in which case 1×10^6 m^3/sec flows upward through each unit area. The upward-bent arrows at the upper right hand corner of each box in Fig. 5 represent this upwelling. The figure associated with these arrows is the approximate total upward flux across the 3000 metre level out of each box. The meridionally directed arrows connecting boxes (such as the one labelled 15 between the two Northern Pacific sub-areas) is meant to represent the interior flow from one sub-area to another. The flow (in 10^6 m^3/sec) between the western boundary currents and each box is indicated by a zonally directed arrow on the left of the box. Thus there is a flux from the western boundary current to the southern sub-area in the Northern Pacific of 20×10^6 m^3/sec and a flow from the northern sub-area of the North Pacific into the western boundary

current of 10×10^6 m^3/sec. As we can see, only half of the water entering the interior of the North Pacific actually flows directly up through the thermocline, the other half recirculates through the western boundary current in accordance with the theory of Part I, section 3. The total flux of 10×10^6 m^3/sec up through the main thermocline in the Northern Pacific is supplied by a western boundary current across the equator, there being no mean flow across the equator at any distance from the western boundary. A similar picture prevails in the North Indian Ocean, but on a much smaller scale, the areas involved being so much smaller than those of the North Pacific. In the North Atlantic the interior flow patterns are much the same, but the strength of the western boundary current is much increased and flows toward the south and across the equator, the recirculation flow being amplified by one of the major deep water sources, S_1. The northern sub-areas of the major southern basins are similar to those just above the equator, but the flow pattern in the southernmost sub-areas (the bottom five in Fig. 5) are quite different. The flow through Drake Passage is not determined in our model so that we must assign it an arbitrary value R, which stands for another kind of recirculation, the amount of water which goes round and round the Antarctic Continent through Drake Passage. The rest of the numbers are obtained from simple continuity requirements. One of the most extraordinary features of this construction is the large flux (30×10^6 m^3/sec) in the western boundary current in the South Pacific, located we suppose, somewhere in the vicinity of the Tonga–Kermadec Trench. Other features of interest, though not so marked, are the reversals in directions of the deep western boundary currents in the Northern Pacific and Indian Oceans. Another feature, which we have also mentioned before (STOMMEL, 1958) is the fact that the cold deep waters flow away from the equator in the interior of the oceans.

Having described Fig. 5, let us now pause to interpret its meaning, and to review its implications. First, we have paid no attention to the probable disturbing effects of pronounced bottom topography within ocean basins. This will doubtless cause deviations from the extremely simple schema we have presented. However, it does not seem worthwhile to try to be more sophisticated than the present stage of our understanding and observational knowledge of the abyssal circulation demands. Similarly, since our main purpose is to imitate the grossest features of the abyssal circulation, we do not pursue the refinements of geographical variation of the upward flux of water to which the theory of the oceanic thermocline would naturally lead us: we simply take a constant and uniform upward velocity over the whole ocean. The pattern of flow in the interior of the deep ocean is thus fixed. Its amplitude remains to be determined. The flow in the interior is independent of the positions of the sources (S_0 and S_1). In a steady state the sum of the sources must be equal to the vertical flux over the entire ocean. Wherever the sources may be we can connect them to the interior solutions as necessary by a system of western boundary currents and the special zonal currents that connect western boundary currents at different boundaries.

We envisage the ultimate cause of the abyssal circulation to be the mechanism of the main thermocline which in effect, as a result of surface heating and applied wind stress, demands a certain upward flux at mid-depth. In the next section we will compute this upward flux from the theory of the oceanic thermocline (ROBINSON and STOMMEL, 1959).

Vol 6 1959

The exact location of the sources is, according to our view, more or less a climatological accident. One could disappear and another appear somewhere else – for example, in the North Pacific – with changing climatic conditions, but these would be accompanied by no major modifications of the interior region of the abyssal circulation, and with no major change of the amplitude of the overall abyssal circulation. Of course any changes in the parametric mixing in the ocean would have effects on the rate of the abyssal circulation, and if the wind convergences were greatly intensified, this might also reduce the overall amplitude of the abyssal transports. This way of looking at the abyssal circulation is quite different from the traditional view which implies that the magnitude of the abyssal circulation is determined essentially by the amount of winter cooling at the polar regions and hence that a warming of polar regions of only a few degrees would largely stop the abyssal flow. From our point of view a warming of polar regions of one or two degrees would not affect deep transports except possibly to shift the location of the sources, and to reshuffle the western boundary currents. The deep temperatures would eventually rise a few degrees, and the distribution of oxygen might change – especially if sources shift position – but the overall transports, and the depth of the thermocline etc. would remain unchanged, providing the climatological factors which determine the thermocline remain unchanged.

Vol. 6 959-60

ESTIMATE OF THE AMPLITUDE OF THE ABYSSAL CIRCULATION ON THE BASIS OF THE THEORY OF THE THERMOCLINE

A dynamical model of the thermocline in principle makes it possible to estimate the vertical component of velocity at mid-depths from the observed distribution of temperature (or more properly density) in the ocean. The calculation is only an estimate because there are simplifications and idealization in the theory which at best are only approximately true in nature and because there appear to be irregularities of a more or less short period observed in hydrographic data which make it impossible to obtain a sufficiently precise mean distribution of properties (in present day language, ' there is a high noise level '), so that one can obtain only a rough estimate of various derivatives of the temperature and salinity fields.

The calculation may be made in a variety of ways.

(1) Pairs of hydrographic stations at the same latitude may be chosen, and a calculation of the depth of no meridional motion, and of the vertical component of velocity, made by the numerical integration of the linearized vorticity equation (STOMMEL 1956, equation 8). To make sure of this method it is necessary to have a good estimate of the mean curve of the wind-stress at the surface, to which unfortunately the results are quite sensitive. If the hydrographic station data contain short period variabilities (internal waves) the calculations of horizontal gradients of density can be made only very crudely. This difficulty has been encountered by most oceanographers employing ' dynamic calculations.' In order to obtain even a fair representation of the currents the pairs of stations must be chosen fairly far apart, or else a line of stations must be smoothed in order to obtain a mean horizontal gradient. The method has one major advantage – it does not require hypothetical statements about relative magnitude of terms in the equation of density transport. Due to the uncertainties introduced through the hydrographic data and by the estimates of

curve of wind-stress the numerical results are doubtful – and it is impossible to state definitely even the range of doubt. STOMMEL's 1956 calculations probably yield vertical velocities that are too high by a factor of ten.

(2) Alternatively, one may make a comparison with an idealized theoretical model such as that of ROBINSON and STOMMEL.

If we start with the same assumptions that ROBINSON and STOMMEL (1959) employ in their theory of the oceanic thermocline, except that we use spherical co-ordinates instead of a beta-plane, we have the following set of physical equations. First the dynamical equations are simplified (see Part I) by the traditional approximations used in spherical co-ordinates for a shallow fluid layer on a rotating globe to the linear geostrophic set :

$$-z\omega a\cos\theta\cdot u = -\partial p/\partial\theta \tag{4.1}$$

$$-z\omega a\cos\theta\cdot v = -\partial p/\partial\phi \tag{4.2}$$

$$g(1-\alpha T) = -\partial p/\partial z \tag{4.3}$$

The equation of continuity is :

$$a\sin\theta\frac{\partial w}{\partial z}+\frac{\partial}{\partial\theta}(u\sin\theta)+\frac{\partial v}{\partial\phi}=0 \tag{4.4}$$

and the equation for the temperature transfer is :

$$w\frac{\partial T}{\partial z}+\frac{u}{a}\frac{\partial T}{\partial\theta}+\frac{v}{a\sin\theta}\frac{\partial T}{\partial\phi}=\kappa\frac{\partial^2 T}{\partial z^2} \tag{4.5}$$

where T is temperature and κ is the parametric mixing coefficient which is discussed at length by FOBINSON and STOMMEL. Equations (4.1) to (4.5) lead to the following set of equations in u, w and T, which in fact are of the same form as the corresponding set in the beta-plane :

$$\frac{\partial^2 w}{\partial z^2}=\left[\frac{g\alpha}{2\omega a^2\cos^2\theta}\right]\frac{\partial T}{\partial\phi} \tag{4.6}$$

$$u\frac{\partial T}{a\,\partial\theta}+w\frac{\partial T}{\partial z}=\kappa\frac{\partial^2 T}{\partial z^2} \tag{4.7}$$

Hence there is no difficulty in immediately applying the same similarity transformation as before. In applying the theory to the real ocean the actual surface temperature cannot be used unless full account is also taken of the wind-stress distribution as well. However, if the origin ($z=0$) is taken at the level where $w=0$ – that is roughly at the inflexion point in the temperature sounding – the temperature boundary condition may be taken as

$$T=T_0\sin^n\theta$$

where of course T_0 is not the amplitude of observed temperature range at the surface, but at the centre of the main thermocline. We obtain the following expression for the vertical component of velocity just beneath the thermocline

$$w(-\infty)=T_0\sin^n\theta\left[\frac{4}{3}\frac{g\kappa^2}{2\omega T_0^2}\frac{\alpha}{a^2\phi}\frac{\sin^{1-2n}\theta}{\cos^2\theta}\right]^{\frac{1}{3}} \tag{4.8}$$

Vol 6 1959

If we could neglect the influence of the wind we would probably set $n = 3$, which would give a pretty good fit to the observed surface temperature distribution in the oceans. However, as shown in ROBINSON and STOMMEL (1959), extensive regions of Ekman convergence produced by the wind at the surface force the level of no vertical component of velocity several hundred metres beneath the surface, and thereby reduce the amplitude of the temperature which should appear in the above expression for the vertical velocity to about one-half the surface amplitude. Instead of $T_0 = 30$ °C we should use $T_0 = 15°$ as an approximation. This has very little influence on the thickness of the thermocline – since it affects it only by a multiplicative factor of the cube root of two, but the amplitude of the vertical component of velocity is directly proportional to this temperature amplitude so that it is reduced to a half of what it would be in the absence of the wind – other factors being equal.

We now wish to perform an integration of the above expression for the vertical component of velocity at mid-depth, over the surface of a single hemisphere globe completely covered with water and bounded by a complete meridian. This is not a bad approximation to the area of the 3000 m. level on the real ocean (about 60 per cent in fact). The limits of integration therefore will be

$$\theta = o, \qquad \theta = \pi/2, \qquad \phi = o, \qquad \phi = 2\pi$$

and the total integral I is :

$$I = \int_0^{\frac{\pi}{2}} \int_0^{2\pi} w\,(-\infty)\, a^2 \sin\theta \, d\theta \, d\phi$$

or

$$\doteq 11 \left[\frac{g\alpha T_0}{\omega} a^4 \kappa^2\right]^{\frac{1}{3}} \tag{4.9}$$

Now setting in the following numerical values used by ROBINSON and STOMMEL :

$$g = 10^3 \text{ cm sec}^{-2} \qquad \alpha = 2 \times 10^{-4} \text{ °C}^{-1}$$

$$\omega = 0{\cdot}7 \times 10^{-4} \text{ sec}^{-1} \qquad \kappa = 0{\cdot}7 \text{ cm}^2 \text{ sec}^{-1}$$

$$a = 0{\cdot}64 \times 10^9 \text{ cm} \qquad T_0 = 15 \text{ °C}$$

we obtain

$$I = 50 \times 10^6 \text{m}^3\text{sec}^{-1}$$

which is in reasonable agreement with the minimum value of $40 \times 10^6 \text{m}^3/\text{sec}$ obtained in the previous section from estimates of flow from the cold sources S_0 and S_1. The average amplitude of the maximum vertical velocity is 4×10^{-5} cm/sec. Since the maximum upward velocity component is at a depth of about 1000 m. on the average, and there is on the average about 3000 m. of water beneath this depth, the deep water is replaced (in a very gross sense of course we have to allow for particular deviations) every 200 years. Although a calculation of this kind is simple because it uses a minimum of data – all the integrations are performed before comparison of the model with observation – it is rather misleading in its simplicity because it contains two implicit assumptions : Namely, that the eddy mixing parameter does not vary geographically (or with depth) and that horizontal mixing and the east-west component of density advection are negligible. These assumptions cannot be fully justified, and

the numerical values of amplitude of the abyssal circulation obtained in the manner outlined above are thus subject to serious uncertainty.

(3) An intermediate approach is therefore indicated. Observational material is used to give an indication of the thickness of the thermocline at a number of different latitudes in the western sides of the interior of each ocean basin. The ROBINSON–STOMMEL theory for infinite depth model is used to compute the deep vertical component of velocity. If there are geographical variations in the mixing parameter they are thus taken into account (the hypotheses that the vertical eddy coefficient is constant in depth, and that horizontal mixing east-west advection are negligible, remain). In order to avoid calculation of too many quantities, zonal averages in each ocean can be calculated by direct x – integration using the analytical form introduced by the similarity transformation. The final summation is a simple numerical integration, where each zonal average of vertical component of velocity is multiplied by an appropriate area, so as to cover the whole ocean; the final sum is the total amplitude of the abyssal circulation. Considering the influence of errors in procedure (1) and of unsubstantiable hypothetical elements in the procedure (2), it seems to us that this intermediate procedure (3) is more likely to give good results, although one can hardly expect an accuracy of better than a factor of two or three.

Thus, using the notation and equation number of ROBINSON and STOMMEL (1959) we want to compute the average of w_∞ – the deep water asymptotic value of the vertical component of velocity along a latitude circle from one side to the other of the interior of a real ocean basin. One of the quantities which we want to avoid having to estimate is w_e, the vertical velocity just beneath the Ekman layer. Between equations (35) and (37) (of ROBINSON and STOMMEL 1959) the similarity transformation W_e of w_e can be eliminated to obtain the following form :

$$W_\infty = -\frac{L^2 \theta_1}{4}$$

where θ_1, is the temperature near the middle of the thermocline, and L is a transformation of thermocline thickness z_t.

$$L = 2\,(\epsilon/x)^{\frac{1}{4}}\, z_t$$

where ϵ is the dimensionless small ratio $(\alpha\beta g)/(3f^2(y)$

Also $$W_\infty = (x/\epsilon)^{\frac{1}{4}}\, w_\infty$$

The calculations for which these formulae hold include the effect of wind-stress at the surface.

The thickness of the thermocline, z_t, is a function of distance from the eastern coast, $-x$ and for a position far in the western side of the interior, $x = x_0$, is measured as $z_t\,(x_0)$. The quantity $z_t\,(x_0)$ is the vertical distance (in centimetres) over which the temperature decreases exponentially to e^{-1} of the mid-thermocline value. By eliminating W_∞ and L, the following expression is obtained :

$$w_\infty = -\epsilon\,[z_t\,(x_0)]^2\,(x - x_0)^{-\frac{1}{2}}\,\theta_1$$

The average of vertical velocity taken along a latitude circle between $x = 0$ and $x = x_0$ is :

$$\overline{w_\infty} = \frac{1}{x_0}\int_{x_0}^{0} w_\infty\, dx$$

The theoretical model uses the condition $z_t = 0$ at $x = 0$ as an eastern boundary condition. In the real ocean z_t is smaller is $x = 0$ than elsewhere but does not actually vanish.

$$\overline{w_\infty} = \frac{3}{2}\frac{1}{x_0}(\epsilon[z_t(x_0)]^2\theta_1)$$

If the depth of the thermocline does not vanish at $x = 0$, the factor 3/2 is replaced by unity.

For the North Pacific Ocean we might take very roughly the calculation as shown in Table 1. Evidently the zonal average of subthermoclinal vertical component of velocity cannot be calculated nearer to the equator than about 10° by the model

Table 1

Lat °*N*	$z_t(x_0)$ (cm)	ϵ (°C^{-1} cm^{-1})	θ_1 (°C)	$\theta_1 \epsilon z^2$	x_0 (cm)	$\overline{w_\infty}$ (cm sec^{-1})
60	6×10^4	$0{\cdot}5 \times 10^{-6}$	5·0	$9{\cdot}0 \times 10^3$	4×10^8	$2{\cdot}5 \times 10^{-5}$
50	5	0·8	5·0	10·0	6	2·0
40	4	1·4	10·0	22·0	8	3·0
30	4	2·4	10·0	38·0	10	3·8
20	3	6·0	10·0	54·0	10	5·4
10	2	24·0	10·0	96·0	10	9·6
0	2	∞	10·0	∞	10	∞

Table 2

METHOD	*Total flux in world abyssal circulation* (10^6 m^3 sec^{-1})	*Vertical velocity under thermocline* (cm day^{-1})	*Replacement time of deep water* (years)
T-S diagram to establish proportions of North and South Atlantic water present; and direct deep current measurements off Charleston to give flux of North Atlantic component	15	0·5	1800
Same T-S analysis and WORTHINGTON's Nova Scotia section and 2000 metre reference level to establish flux of North Atlantic Component	40	1·3	650
Same T-S analysis and WÜST's dynamic calculations for South Atlantic Ocean for flux of North Atlantic component	50	1·6	550
Depth of the thermocline over the world ocean and theory of thermocline to establish total flux	90	3·0	300

proposed by ROBINSON and STOMMEL. This is obvious because the parameter ϵ is not small for latitudes less than 10° and because such observed features of the real ocean circulation as the Cromwell Current (KNAUSS 1959) are not shown when the theory is applied is this inadmissable range. If we can assume that the deep vertical velocity at the equator is not in some way quite singular we can compute averages

and obtain thus a figure of abour 4×10^{-5} cm sec^{-1} as a North Pacific average (excluding a band of latitude south of 10 °N). This figure may be off by a factor of three, but at least the estimate of the mean wind-stress and the errors due to internal waves and details of hydrographic structure do not enter.

In a letter to the editor of *Deep-Sea Research*, STOMMEL (1958) gave a value of 3×10^{-5} cm sec^{-1} based on a much cruder thermo-convective model. Both of these are perhaps in fact too large by a factor of two or three. A figure more nearly in accord with WÜST's (1957) geostrophic calculation is $1{\cdot}5 \times 10^{-5}$ cm sec^{-1}. A smaller value than this is obtained if we assume that the transport in the deep western boundary current in the North Atlantic is always the same as that measured in March–April 1957 by the joint British–American expedition (SWALLOW – WORTHINGTON 1957) 6×10^{12} cm^3/sec. Since the T-S properties of abyssal water (Section 3) indicate that the total flux of the abyssal circulation is from 2 to 3 times that of the western boundary current in the N Atlantic, we might choose 14×10^{12} cm^3/sec as the appropriate figure. To obtain the world average vertical velocity we divide this figure by the area of the ocean at 2000 m. (3×10^{18} cm^2) and obtain a figure of: $0{\cdot}5 \times 10^{-5}$ cm sec^{-1}. Table 2 summarizes these calculations.

EFFECT OF LATERAL MIXING ON THE DYNAMICS OF FLOW

In the preceding sections we have proposed a very highly idealized model of the abyssal circulation. The theoretically predicted flow patterns are based on the fundamental assumptions that (a) flow in the interior of the ocean is geostrophically balanced with negligible *dynamical* influence from frictional or eddy effects (b) significant departure from geostrophy occurs only in boundary currents along the edges of continental masses. Since it is clear that both lateral and vertical eddy diffusivity play a very important role in many oceanic processes and in the distribution of tracer properties of the water masses, it is important to inquire, at least in terms of orders of magnitude, as to whether Austausch coefficients large enough to be important in determining the distribution of oxygen, radio carbon, salinity, etc. are at the same time large enough to affect the dynamics of the postulated planetary motions driven by source-sink distributions.

To answer this question it is appropriate to assess the order of magnitude of the various terms appearing in the dynamical equations and in the diffusion equation.

Consider the dynamical problem in the form expressed by the vorticity equation for steady two-dimensional flow on a β-plane, retaining lateral friction terms :

$$u\frac{\partial \eta}{\partial x} + v\frac{\partial \eta}{\partial y} + \eta \,\mathrm{div}_{\mathbf{H}}\,\mathbf{v} + f\,\mathrm{div}_{\mathbf{H}}\,\mathbf{v} + \beta v = A_H\left[\frac{\partial^2 \eta}{\partial x^2} + \frac{\partial^2 \eta}{\partial y^2}\right]$$

In this notation :

x, y co-ordinates positive toward the east and north respectively,

u, v particle velocities corresponding to co-ordinates x, y respectively.

$\mathbf{v}$, total vector velocity

$$\eta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$$

$\mathrm{div}_{\mathbf{H}}$, horizontal divergence

f, Coriolis parameter

β, $\partial f/\partial y$

A_H, horizontal (lateral) eddy coefficient.

Transforming Equations 3·1 of Part I to the β-plane as defined above, the zonal and meridional components of geostrophically balanced flow for a uniformly distributed sink Q_0 over the entire surface are :

$$u = 2Q_0 (x_2 - x)/h \tag{2.2}$$

$$v = fQ_0/\beta h \tag{2.3}$$

$$\eta \doteqdot - \frac{2Q_0}{h} \frac{(x_2 - x)}{a} = - \frac{u}{a} \tag{2.4}$$

The order of magnitude of Q_0, estimated in section 4 is 4×10^{-5} cm/sec. We adopt the following values for various relevant quantities :

Horizontal distance scale : $X \sim Y \sim 6000 \text{ km} = 6 \times 10^8 \text{ cm}$

Thickness (depth) of abyssal layer : $h \sim 3 \text{ km} = 3 \times 10^5 \text{ cm}$

Coriolis parameter : $f \sim 10^{-4} \text{ sec}^{-1}$

$\beta \sim 2 \times 10^{-13} \text{ sec}^{-1} \text{ cm}^{-1}$

Radius of the earth : $\sim 6{\cdot}5 \times 10^8 \text{ cm}$

Vertical velocity at top of abyssal layer : $Q_0 \sim 4 \times 10^{-5} \text{cm/sec}$

From these assumed values we estimate the order of magnitude of the following :

$$u \sim 0{\cdot}16 \text{ cm/sec}$$

$$v \sim 0{\cdot}067 \text{ cm/sec}$$

$$\eta \sim + 2{\cdot}5 \times 10^{-10} \text{ sec}^{-1}$$

$$\text{div}_H \, \mathbf{v} \sim - 1{\cdot}3 \times 10^{-10} \text{ sec}^{-1}$$

$$\partial\eta/\partial x \sim - 4{\cdot}1 \times 10^{-19} \text{ sec}^{-1} \text{ cm}^{-1}$$

$$\partial\eta/\partial y \sim + 3{\cdot}8 \times 10^{-19} \text{ sec}^{3-1} \text{ cm}^{-1}$$

$$\frac{\partial^2 \eta}{\partial x} \sim \frac{\partial^2 \eta}{\partial y^2} \sim 7 \times 10^{-28} \text{ sec}^{-1} \text{ cm}^{-2}$$

(The magnitudes of the derivatives have been estimated by finding the appropriate derivates of Equations 3.1, Part I and then converting to the β-plane).

Thus we have the following orders of magnitude for the terms in Equation (2.1).

$$u \frac{\partial \eta}{\partial x} \sim - 7 \times 10^{-2}{}_0 \text{ sec}^{-2}$$

$$v \frac{\partial \eta}{\partial y} \sim 3 \times 10^{-20} \text{ sec}^{-2}$$

$$\eta \, \text{div}_H \, \mathbf{v} \sim - 3 \times 10^{-20} \text{ sec}^{-2}$$

$$f \operatorname{div}_H \mathbf{v} \sim -1{\cdot}3 \times 10^{-14} \text{ sec}^{-2}$$

$$\beta v \sim +1{\cdot}3 \times 10^{-11} \text{ sec}^{-2}$$

It is clear that the dominant terms on the left-hand side of Equation (2.1) are $f \operatorname{div}_H \mathbf{v}$ and βv. It is, of course, the balance of these terms which represents the situation we describe as 'purely geostrophic flow.' We can now find values of the Austausch coefficient A_H which would make the right hand side of the equation assume magnitudes comparable with the dominant terms on the left. Noting that the latter are of the order of 10^{-14} and that the second derivatives on the right are of the order of 10^{-27}, it is apparent that A_H, must be of the order of 10^{12} cm^2/sec before the frictional terms on the right attain the order of one tenth of the planetary divergence terms.

Available estimates of the large scale lateral eddy coefficient for the ocean usually fall in the range 10^6 to 10^8 cm^2/sec. In the light of these results we feel that our fundamental assumption, that the very low velocity flows induced in the interior of the ocean are essentially geostrophically balanced, is a reasonable one. We not do expect lateral friction to affect seriously the purely geostrophic dynamical foundation of the basic circulation patterns. (It is interesting to note that BURGER (1958) obtains a similar result for large scale flows in his consideration of planetary motions in the atmosphere).

We now make a similar scale and order of magnitude analysis of diffusion of a tracer property of the water. The basic differential equation is of the form :

$$A_H \nabla^2_H c - \mathbf{v} \cdot \nabla_H c - \lambda c = 0 \tag{2.5}$$

where c denotes concentration of a property (say oxygen) and λ is the reciprocal of the " half life " of decay. The three terms of Equation (2.5) represent a balance among diffusive transport, advective transport, and decay, the amplitude of the advective term being determined by the basic geostrophic flow.

To examine the relative significance at the planetary scale of the three terms in Equation (2.5) we re-write it in the following 'operator' form :

$$\left[A_H \nabla^2_H - \mathbf{v} \cdot \nabla_H - \frac{3{\cdot}17 \times 10^{-8}}{T} \right] c = 0 \tag{2.6}$$

where T is the half-life of decay in years of the tracer property. Introducing the scale magnitudes used earlier, the orders of magnitude of the terms appearing in (2.6) are indicated by :

$$\left[2{\cdot}8\,(10^{-18})\, A_H - 2{\cdot}7\,(10^{-10}) - \frac{3{\cdot}17 \times 10^{-8}}{T} \right] c = 0 \tag{2.7}$$

All three terms are of roughly comparable significance when A_H is of the order of $10^7 - 10^8$ cm^2/sec and T is of the order of $10^2 - 10^3$ years. If the half life becomes much less than 10^2 years, the distribution of the property is determined principally by diffusion in the immediate neighbourhood of the western boundary current, advective effects are insignificant, and appreciable concentrations do not extend very far into the interior of the ocean. If T and A_H are both large, the property becomes essentially uniformly distributed, the advective effects insignificant, and the concentration level determined by the relation between half life and rate of supply.

Vol 6 1959

The amplitude of the basic geostrophic circulation in the model we have postulated is such that Austausch coefficients of the order of $10^7 - 10^8$ cm²/sec will have a very marked diffusive effect on the distribution of tracer properties of the water masses. At the same time, as shown earlier in this section, coefficients of this magnitude would be expected to have a *negligible* effect on the dynamics of the basic flow. This rather paradoxical result arises from the very large scale and low velocity of the basic flow, the frictional force terms being proportional to third spatial derivatives of the geostrophic velocities. It is felt therefore that the fundamental model, in which an interior, large scale, geostrophically balanced circulation is driven by the mechanism of the thermocline, is basically reasonable. It is simultaneously to be pointed out that great caution must be exercised in drawing inferences about the basic flow from observed distributions of water mass properties; the latter probably being seriously affected by diffusion on a scale substantially smaller than that of the basic flow.

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Contribution No. 1055 *from the Woods Hole Oceanographic Institution. Part of this work was done while one of us (A.B.A.) was the recipient of a Fellowship of the John Simon Guggenheim Memorial Foundation. The rest was done under contract with the Office of Naval Research.*

Vol. 6 959-60

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SHORTER COMMUNICATIONS

Hermaphroditism in Archibenthic and Pelagic Fishes of the Order Iniomi

GILES W. MEAD

(*Received* 14 *June* 1959)

Abstract—Ovotestes, with both male and female components in advanced stages of maturation, were found in ten species of the iniomous familes Chlorophthalmidae, Bathypteroidae, Alepisauridae and Paralepididae. Both male and female elements were apparently normal, and several species contained sperm-filled vas deferens and mature but unfertilized ova. There is insufficient material of these uncommon deep-sea fishes to provide satisfactory data on possible protandry, proterogeny, or the seasonal reproductive cycle.

Hermaphroditism and sex reversal are well known normal occurrences in certain eels, symbranchids, in the percoid families Serranidae, Sparidae and Maenidae, and ovotestes are not infrequently found terratologically in many fishes irrespective of relationship. A case of voluntary self-fertilization in fishes was recently reported by CLARK (1959 : 215) in the Floridian serranid *Serranellus subligarius*. Certain of the Iniomi must now be added to the list of normally monoecious species, for although these deep-sea forms cannot yet be studied alive, sections of their gonads reveal ovotestes similar in general structure to those of the monoecious percoids. An ovotestis has been found in the following specimens :

Chlorophthalmus agassizi Bonaparte (Chlorophthalmidae, benthonic).

An ovotestis, with neither component mature, was found in an adult specimen 137 mm in standard length which was caught by the U.S. Fish and Wildlife Service exploratory vessel *Oregon* at station 1054, Campeche Bank (Gulf of Mexico), at a depth of 200 fm

Chlorophthalmus brasiliensis Mead. Ovotestes with mature sperm and ova which were nearly or quite mature were found in the two individuals examined. These fishes, about 120 mm in length and representative of the total catch, were taken at *Oregon* stn. 2082, off equatorial Brazil at a depth of 200 fm.

Parasudis truculentus (Goode and Bean) (Chlorophthalmidae, benthonic). Ovotestes which were not fully mature were found in four individuals, each about 130 mm long, which were caught at *Oregon* stn. 2005, off Surinam at 200 fm (Fig. 1). These four were smaller than most of those caught at this locality. The ovotestis of a more typical individual, about 160 mm long, contained mature ova in both the ovary and the oviduct, sperm-filled ducts, and a small testicular area medial to the ovarian component of each gonad.

Benthosaurus grallator Goode and Bean (Bathypteroidae, benthonic). A large individual (290 mm, the largest specimen known from the western Atlantic), from *Oregon* stn. 1303, in the Gulf of Mexico south of Mississippi at 1150 to 1200 fm, contained a mature ovotestis and a sperm-filled vas deferens.

Bathypterois (*Hemipterois*) *viridensis* (Roule) (Bathypteroidae, benthonic). A well developed and mature ovotestis was found in one of the largest known individuals, a 180 mm fish from *Oregon* stn. 1914, in the western Carbibbean at 350 fm. A smaller individual, 121·0 mm long, from *Oregon* stn. 1907, in the western Caribbean at 400 to 425 fm, contained an immature ovotestis.

Bathypterois (*Bathypterois*) *quadrifilis* Günther. An ovotestis was found in three of the five individuals studied, 121 to 142 mm in length, from *Oregon* stn. 1426, in the Gulf of Mexico south of Alabama at 600 fm. The other two appeared to have ovaries only, although the entire gonad was not sectioned. The sex products were mature, and there was no correlation between the occurrence of an ovotestis or an ovary with size.

Vol
6
1959

Vol.
6
959-60

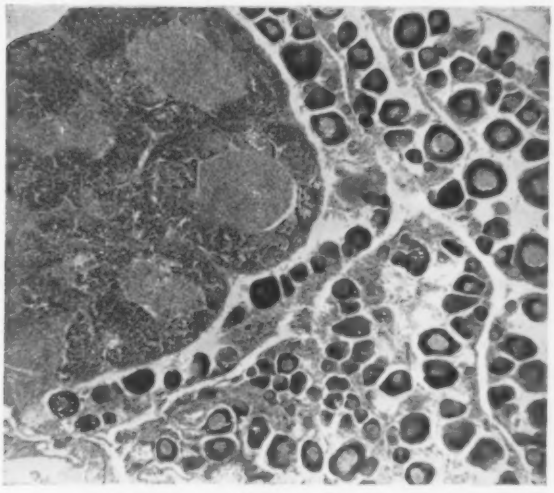

FIG. 1. Part of the ovotestis of *Parasudis truculentus*.

Vol
6
1959

Bathypterois (*Bathypterois*) *longipes* Günther. The only available specimen, a 150 mm adult in poor condition taken by the *Atlantis* off the Azores in August of 1947 at 3200 fm, has an ovotestis. The degree of maturity could not be determined.

Bathypterois (*Bathypterois*) *bigelowi* Mead. An ovotestis, neither component of which was mature, was found in a 117·5 mm individual from *Oregon* stn. 1917, in the western Caribbean at 325 fm.

Alepisaurus ferox Lowe (Alepisauridae, pelagic). The two individuals studied, one 605 mm long from the Gulf of Mexico, the other 735 mm long from off New England and caught on tuna long-lines by the Fish and Wildlife Service vessels *Oregon* and *Delaware*, contained immature ovotestes.

Lestidium (*Lestidium*) *pseudosphyraenoides* (Ege) (Paralepididae, pelagic; this identification is provisional, the specimen is in the collection of the Marine Laboratory, University of Miami; UMML 3903). One specimen was examined, a 160 mm fish caught at *Oregon* stn. 1944, in the western Caribbean at 275 fm. The ovarian part of the ovotestis was completely spawned out. Fully mature sperm was present in the vas deferens as well as in the testis.

Of the remaining groups of benthonic Iniomi, the Aulopidae and Synodontidae are dioecious and separate sexes have been reported in the abyssal Bathysauridae (GÜNTHER, 1887 : 183). No adequately preserved representatives of the Ipnopidae, Harpadontidae or Scopelosauridae are currently available.

Among the pelagic members of the order, suitable specimens of most Paralepididae and the Scopelarchidae, Omosudidae, Evermanellidae, Anotopteridae are not available; the lantern fishes (Myctophidae and Neoscopelidae) are, as far as known, dioecious. External sexual dimorphism is found in all aulopids and in some myctophids. Hermaphroditism is thus not confined to any taxonomically-related group within the order, or to the iniomous fishes of any given habitat. The alepisauroids (*Alepisaurus*, *Lestidium* and allied forms) are pelagic and are not closely related to the remaining families, such as the Bathypteroidae, which are benthonic.

In all of the monoecious fishes examined, the testicular part of the gonad lies dorso-medial to the ovarian, and the associated sperm ducts course posteriorly in close association with the urinary duct. The small tailless sperm reach the exterior through an opening in or near the tip of a small urogenital papilla. The ova are extruded through an ovopore which is located between the anus and the papilla.

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U.S. Fish and Wildlife Service
U.S. National Museum, Washington, D.C.

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The seasonal and geographical range of primary production in the Western Sargasso Sea

JOHN H. RYTHER and DAVID W. MENZEL

(*Received* 3 *September*, 1959)

IN a recent publication (MENZEL and RYTHER, in press) the annual cycle of primary production was discussed for the North-western Sargasso Sea as observed from a single hydrographic station (Station 'S') located 15 miles SE of Bermuda in 1500 fm of water. The values for net daily organic production

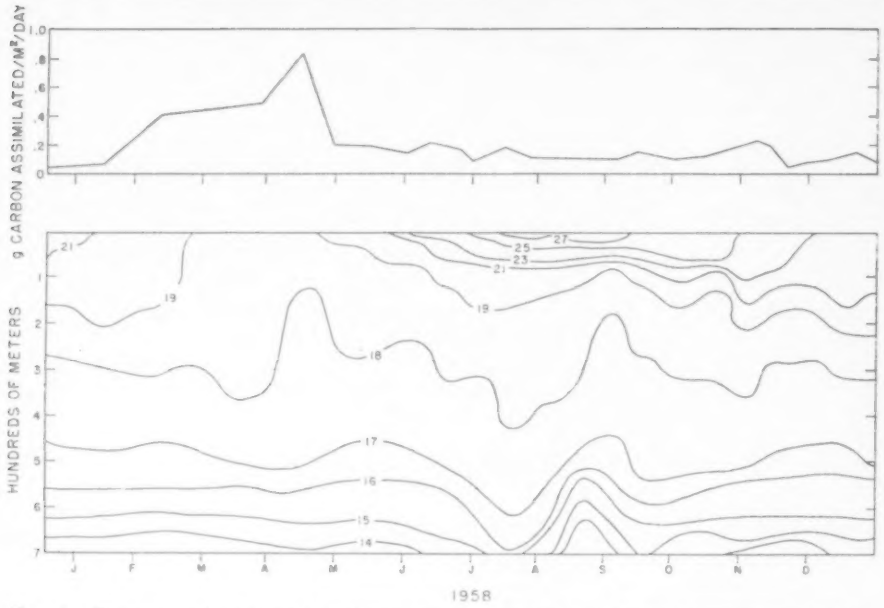

FIG. 1. The seasonal cycle of net primary production and the seasonal temperature profile at Station S for 1958.

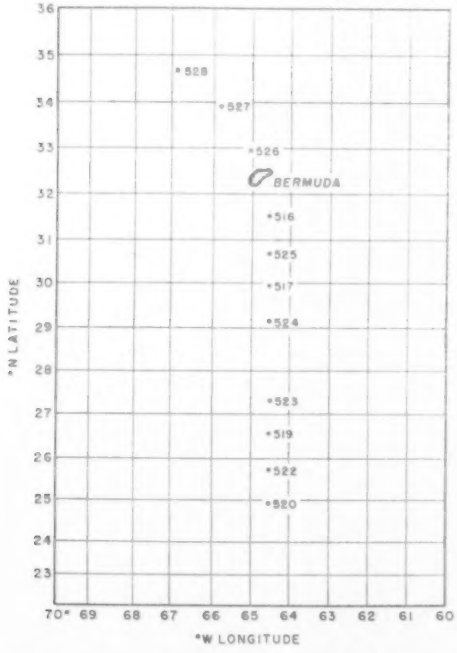

FIG. 2. Station locations for the oceanographic section south of Bermuda made during *Crawford* cruise No. 24 in March, 1959.

Vol 6 1959

(g carbon assimilated/m²/day) for 1958 are shown in Fig. 1. These values were obtained by *in situ* or simulated *in situ* ^{14}C measurements. The latter consisted of exposing samples, collected from several depths within the euphotic zone and inoculated with $^{14}CO_3^=$, to light intensities comparable to those occurring at the depths from which the samples were taken. This was accomplished by means of neutral density filters covering a water-cooled incubator which was mounted on the deck of the oceanographic vessel. The *in situ* measurements lasted for half the daylight portion of a day; the modified *in situ* measurements, for 24 hrs.

There was a definite seasonal cycle of production, comparable to but of less magnitude than the cycle in temperate and boreal waters. The values ranged from a maximum of 0·85 g C/m²/day in April to 0·05 g C/m²/day in November. Also shown in Fig. 1 is the seasonal temperature profile to 700 m, which illustrates the relationship between vertical stability of the surface waters and productivity. Minimum levels of production occurred throughout the summer and early fall when the seasonal thermocline was well developed. Moderately high levels persisted throughout the late fall, winter, and early spring when the water was more or less isothermal and mixed to or nearly to the depth of the permanent thermocline (about 400 m). The highest values for the year occurred briefly in April when thermal stratification of the surface layers first began to develop.

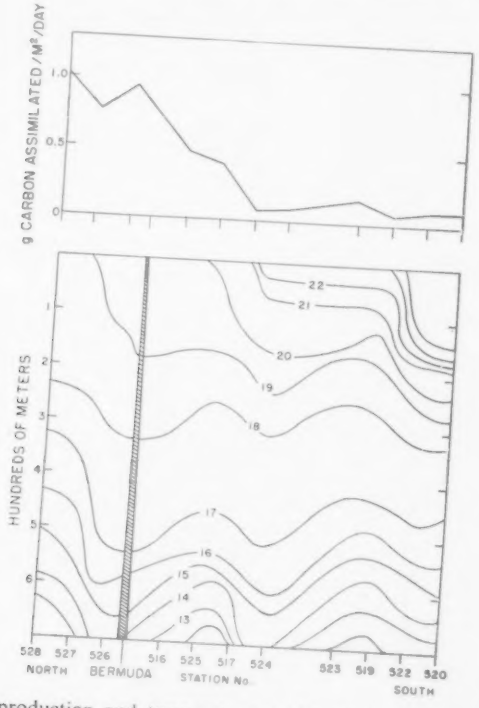

FIG. 3. Net primary production and temperature profile observed on the section illustrated in Fig. 2.

The relatively high production throughout the winter is in contrast to the situation further north, where in deep water there are both winter and summer minima (RYTHER and YENTSCH, 1958). This is due to the conditions in the Bermuda region of (1) greater incident radiation in winter, (2) more transparent water and (3) less extensive vertical mixing. As a result of this combination of factors, the ratio euphotic zone/mixed zone is greater at Bermuda than further north permitting plant growth in winter in the former but not the latter region.

The latitude of Bermuda may be considered a semi-tropical area where surface water conditions differ markedly from those further south in the true tropics. There, no winter cooling and mixing

of the surface layers occurs and the 'seasonal thermocline' of more northern waters persists in the tropics throughout the year and merges more or less continuously into the permanent thermocline. The wide band of 18° water, characteristic of the Sargasso Sea (WORTHINGTON, 1959), becomes thinner and thinner southward from Bermuda as the westerly drifting equatorial current is approached. Just 100 miles south of the island, surface temperatures seldom, if ever, drop below 20 °C.

Since high production is so obviously dependent upon winter mixing and its enrichment of the surface layers (60 per cent of the annual production at Bermuda occurred between January and April, 1958) it follows that the productivity of tropical waters where no such mixing occurs must be low throughout the year. There has been no study made of the annual cycle of primary production in the tropics comparable to that described above for the semi-tropical Sargasso Sea. But the results of a short oceanographic section north and south of Bermuda do much to confirm the hypothesis presented above.

This section, made during the first week of March, 1959 on Cruise No. 24 of the Research Vessel *Crawford*, consisted of two stations north-west of Bermuda and nine stations approximately 50 miles apart in a line due south of the island extending to 24° 53′N latitude. The location of these stations is shown in Fig. 2. It was intended to extend this section across the Sargasso Sea to the north-eastern coast of the United States, but a severe winter storm prevented further working during the passage.

Winter cooling and mixing in the North Atlantic ocean are usually most extreme during early March. Actually, the winter of 1959 was rather mild and the season somewhat early. The first indications of seasonal thermal stratification of the surface water had begun to appear in the Bermuda region by March 1.

Fig. 3 shows the values for net primary production, as measured by the simulated *in situ* ^{14}C technique, on this section. Below this is shown the temperature profile to 700 m. It is remarkably similar to the seasonal cycle at Bermuda as shown in Fig. 1. Production at the northern end of the section averaged about 1·0 g C/m²/day, the highest ever observed in the Sargasso Sea and about comparable to the seasonal maximum at Station *S* in 1958. Production dropped off rapidly south of Bermuda reaching a low, rather constant level at the six southernmost stations of the section of about 0·1 g C/m²/day. Across a distance of less than 200 miles production differed by an order of magnitude.

So, too, does the north-south temperature profile mirror the seasonal trend at Station *S*. The broad band of isothermal water, some 300 m deep, at the northern end of the section merges into a well stratified water mass to the south much the same as the development in time of the seasonal thermocline *in situ* at Bermuda.

No other time of year and no other set of conditions can be envisaged which would cause primary production in the tropical Sargasso Sea to exceed the level shown here. Until a seasonal study of the region has been made, it may be assumed that the process proceeds at a low rather constant level, with an annual rate of some 25–50 gC/m²/year (net production), perhaps one third of that in the semi-tropical waters to the North.

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Vol 6 1959

INSTRUMENTAL NOTE

The contouring temperature recorder

WILLIAM S. RICHARDSON and CHARLES J. HUBBARD

(*Received* 29 *April* 1959)

(1) INTRODUCTION

THE bathythermograph (BT) has been the standard instrument for the study of the thermal structure of the upper layers of the ocean for roughly 25 years. This very useful instrument provides a trace of temperature against depth (to either 450 or 900 ft, depending upon the type used) while the ship is under way. For some purposes this trace is the end result of the experiment, but to study the variations of structure in the horizontal, one commonly reads the BT traces, picks off the depths of various temperatures, and plots these on a curve of depth versus distance. The isotherms are then contoured and the resulting distribution is used as the data from which the physical processes are inferred. It is possible on a very arduous schedule to obtain BT's about every five minutes, but on most trips one is taken every hour or more. This provides sufficient data for many problems, and it may be easily supplemented by continuous surface temperature records to provide an indication of the location of some of the pronounced differences which often occur from one BT to the next. For many purposes the BT data even at its most rapid rate of acquisition is not sufficient; problems in sound transmission, internal waves and microstructure are examples.

The Contouring Temperature Recorder to be described in this paper is a device which takes data from temperature sensors (thermistors) located in a chain towed behind a ship and plots on a continuous record the vertical distribution of isotherms. The thermistors are electronically scanned in sequence from the top to the bottom of the chain, in a time interval adjustable from 2 to 20 secs. If it is desired to do so, the scan rate can be adjusted in relation to the speed of the ship so that the information is taken from a vertical column in the water. In the design to be described, the 1° isotherms from −2° to 32°C can be contoured along with all 0·1° and 0·05°C isotherms between these limits. Also, selection can be made so only the 1°, the 0·1° (which of course include the 1°), or the 0·05°C are contoured. It is possible to accentuate the 1°C isotherms with respect to the others to aid in their identification. Provision is made for determining the depth of the tow (which is a function of ship's speed) and also for selecting the fastest possible scan rate consistent with the complexity of the water column. A straightforward method of isotherm identification is provided; this can be quite a difficult problem if there are many inversions within the water column.

(2) THE HOIST AND CHAIN–MOUNTING OF THERMISTORS AND DEPTH METER

The chain on which the thermistors are mounted is manufactured by Commercial Engineering Company, of Houston, Texas, (Fig. 1). The hoist is diesel-hydraulic powered and carries six hundred feet of chain. The chain consists of links about eleven inches long having a U-shaped cross section in the hotizontal plane. A rubber boot is contained inside the links into which a multi conductor electrical harness can be fitted. The rear portion of each link is then closed with a plastic fairing which holds the wire securely in place and completes the streamlining of the link. A 3400 lb ' fish ' is used on the end of the chain; the design of the hoist includes a simple means of launching and retrieving the fish.

The towing characteristics of the chain are shown in Fig. 2. The tow is very steady at speeds up to at least 15 knots and is normally used at about 10 knots to provide measurements to about 400 ft.

The electrical harness consists of about 100 leads of A.W.G. 26 hookup wire. This provides two leads to each of 23 thermistors, 3 leads to the depth meter and about 50 spare leads for other

uses or repairs. The thermistors are mounted 25 ft apart in holes drilled in the plastic fairings and potted in place with an epoxy resin to insulate the electrical connection from the water. The depth meter, which is a commercial Bourdon type pressure potentiometer with 200 PSI full scale, is mounted in a stainless steel pressure case attached to a fairing 450 ft below the uppermost thermistor.

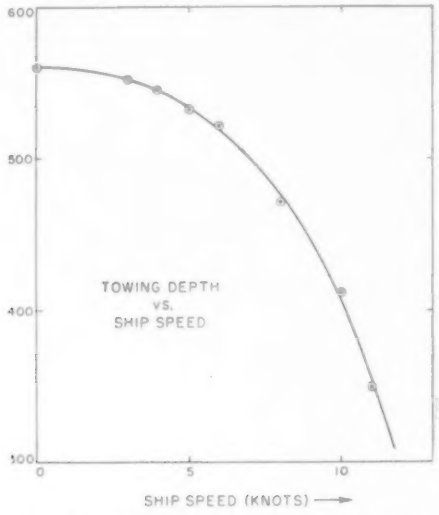

FIG. 2. Towing depth versus ship's speed.

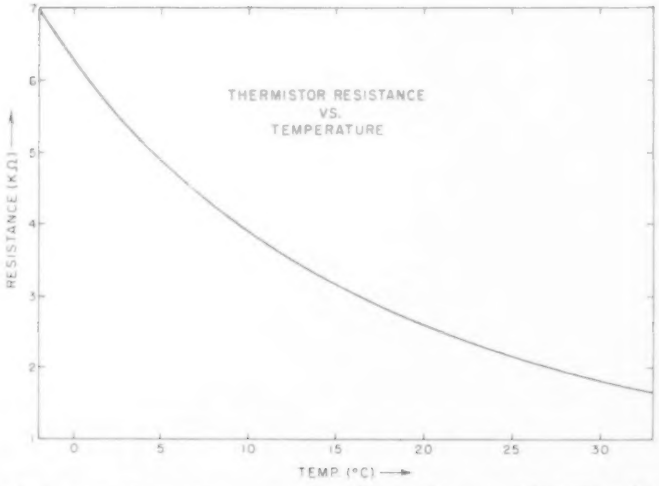

FIG. 3. Typical curve of temperature versus resistance for type 32A1 Thermistor.

(3) THE THERMISTORS

Type 32A1 Thermistors manufactured by Victory Engineering Company, of Union, New Jersey, are used in this equipment. Fig. 3 shows a typical curve of resistance versus temperature for these units. It is necessary that all units used are nearly identical. This condition is met as carefully as possible by procuring a large stock of thermistors from a single bead mix and culling them at a fixed temperature to find a set with the same resistance at this temperature. The set is then checked at

Vol.
6
959-60

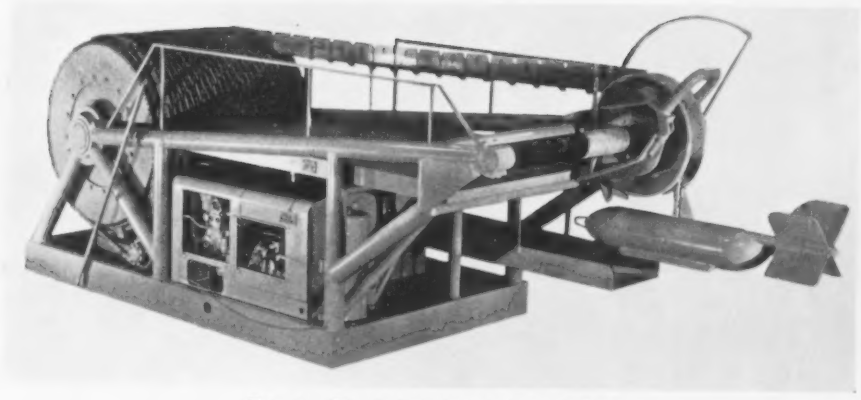

FIG. 1. The hoist with chain installed

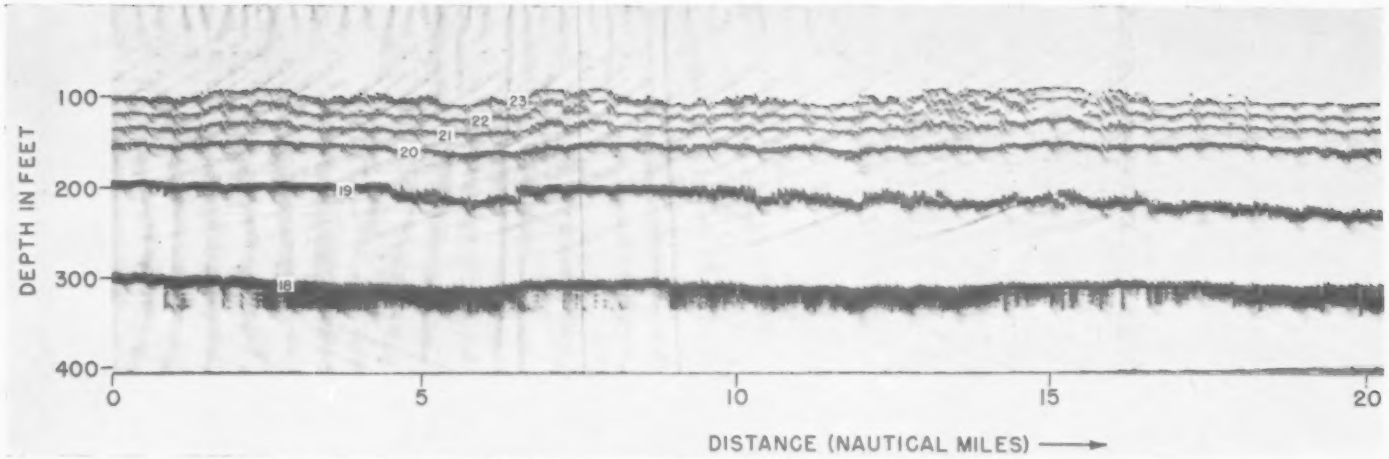

FIG. 7. Vicinity of 39°N 71°W ; 1 October 1958 ; Course 180°T ; Speed 10 knots ; 1°C isotherms.

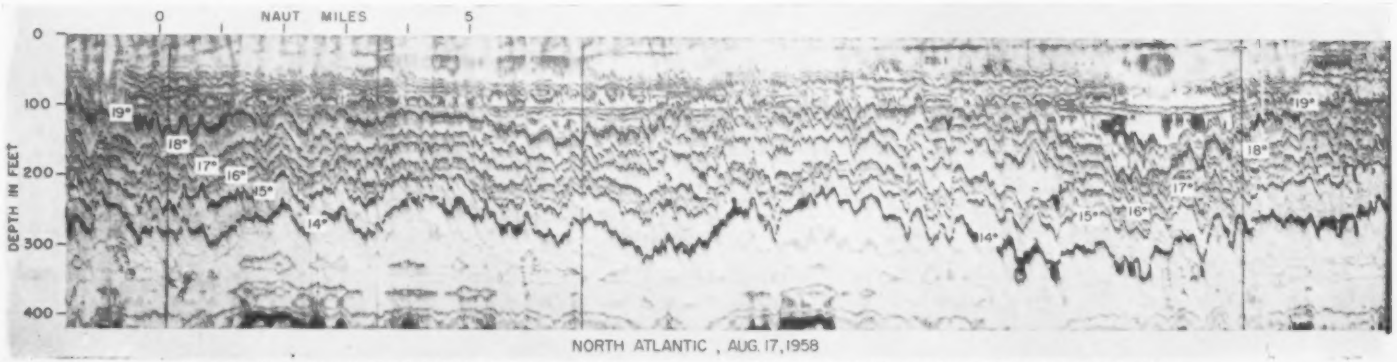

FIG. 8. Vicinity of 39°N 11°W ; 17 August 1958 ; Course 270°T ; Speed 10 knots ; 1°C and 0·1°C isotherms.

1959
6
VOL

Vol.
6
1959-60

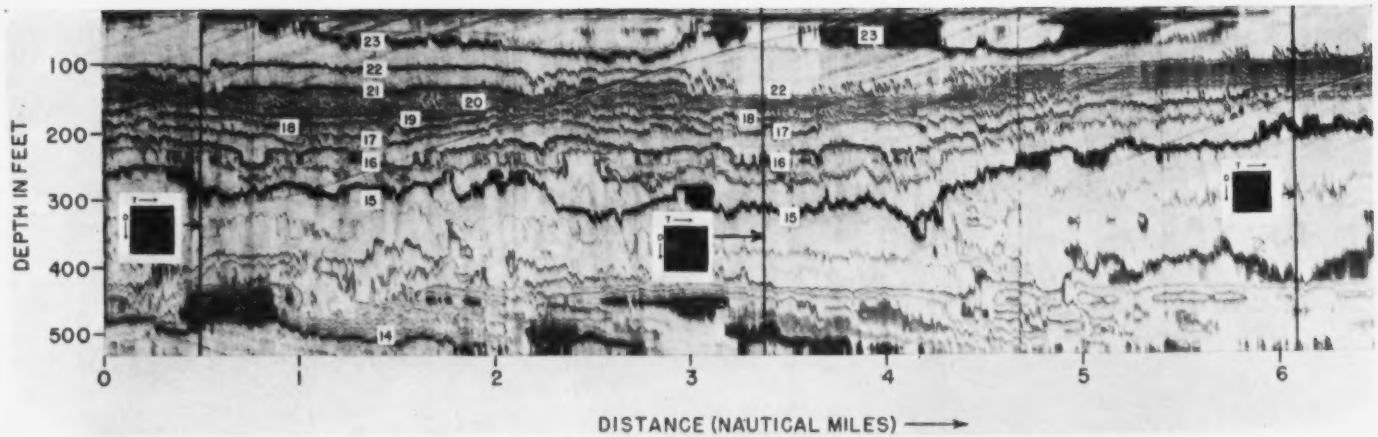

FIG. 9. Vicinity of 6°E 38°N ; 15 July 1958 ; Course 100°T ; Speed 3·3 knots ; 1°C and 0·1°C isotherms.

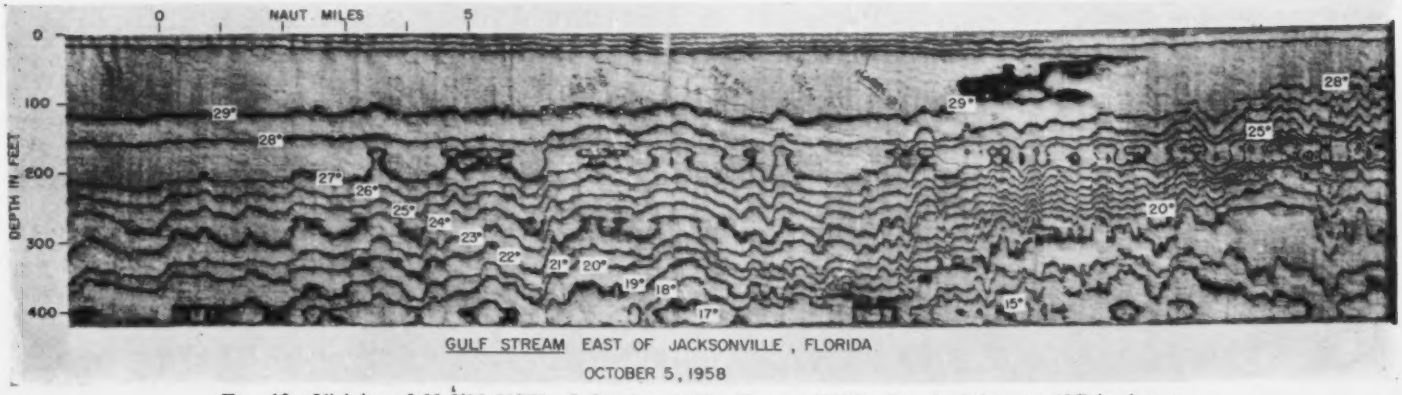

FIG. 10. Vicinity of 30·5°N 81°W ; 5 October 1958 ; Course 270°T ; Speed 10 knots ; 1°C isotherms.

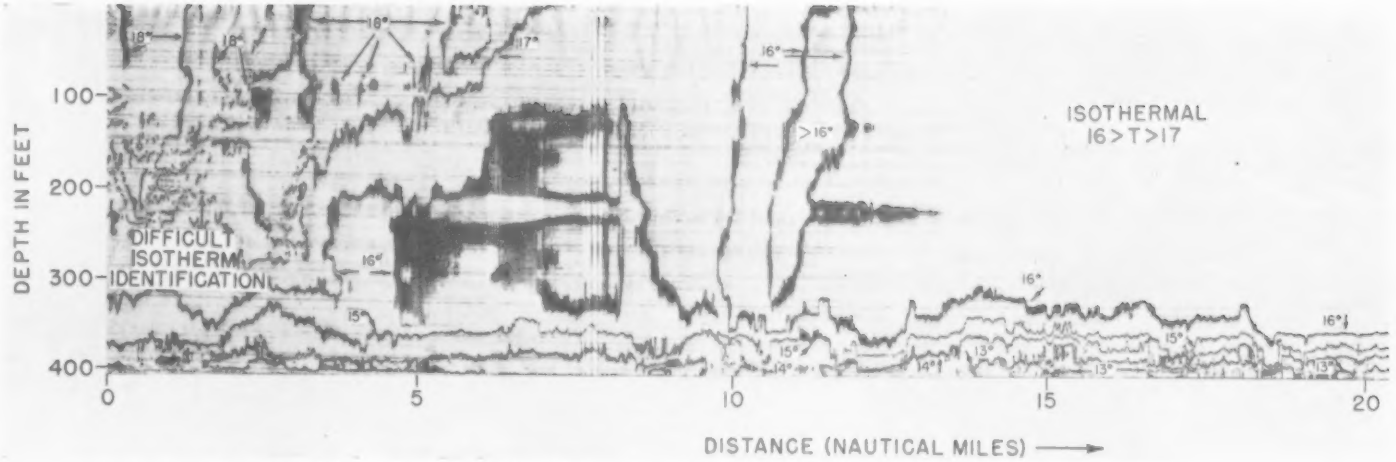

FIG. 11. Vicinity of 37°N 70°W ; 19 December 1958 ; Course 340°T ; Speed 10 knots ; 1°C isotherms.

VOL
6
1959

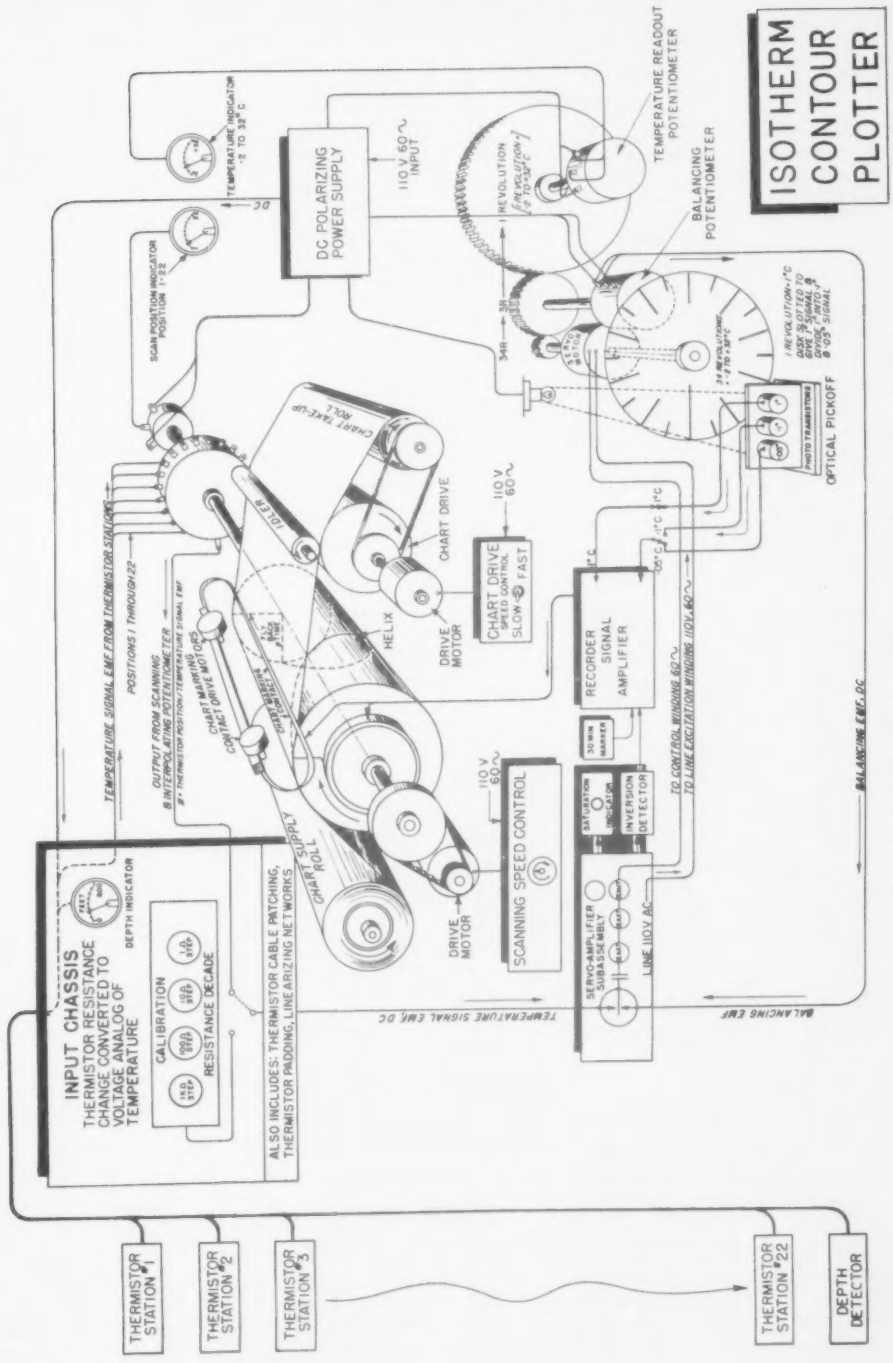

FIG. 4. System drawing of Contouring Temperature Recorder.

another temperature to determine that the curvatures are sufficiently identical and finally selected numbers of the set are calibrated against a Bureau of Standards thermometer calibrated to 0·01°C.

(4) GENERAL DESCRIPTION OF THE SYSTEM

Fig. 4 is a system drawing of the electronics, slightly simplified for clarity at certain points. The operation is as follows. Each thermistor is connected in a circuit like that shown in Fig. 5. The fixed resistance R_W is the lead resistance in the cable to the thermistor. Since this varies with the position of the thermistor in the chain, a second resistor R_P called padding is provided in each circuit.

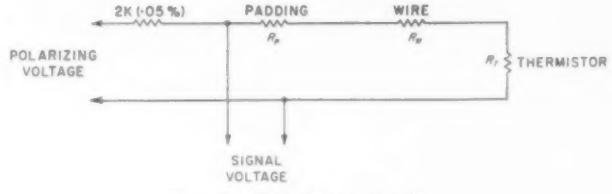

FIG. 5. Thermistor circuit.

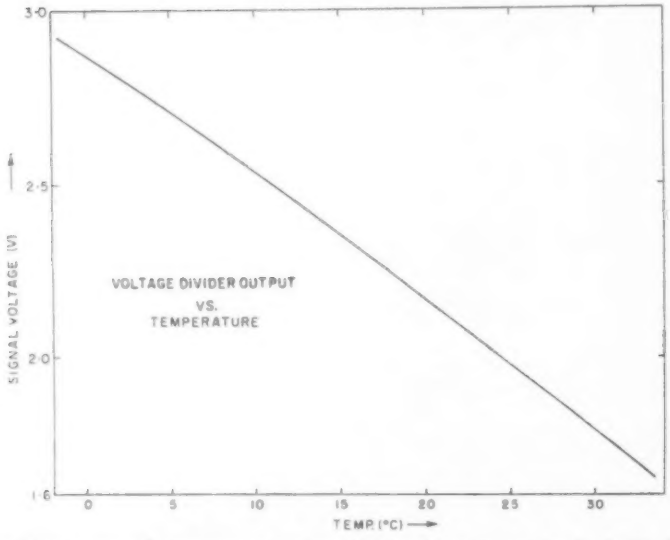

FIG. 6. Typical curve of temperature versus signal voltage from circuit shown in Fig. 5.

These are adjusted so that the sum of R_T, R_W, and R_P has the same value for each circuit at a selected temperature. All these circuits are powered by the same voltage source. These voltage dividers provide a signal output which is very nearly linear with temperature (Fig. 6), and there is one such voltage for each depth where a thermistor is located.

The voltage output from each of these dividers is cabled to a large potentiometer (interpolating and scanning potentiometer) where they are connected to equally spaced taps on the potentiometer winding in the order of their depths. The potentiometer has eighty thousand ohms between adjacent taps (total resistance 2·4 megohms). Therefore, each thermistor-voltage divider looks into an apparently open circuit, and the voltages impressed at the taps of the potentiometer remain proportional to temperature. As this potentiometer is rotated, the voltage on the slider is proportional to the temperature at the depth corresponding to its angle of rotation. When the slider is between taps, the voltage on the slider is a linear interpolation between the voltages impressed on the adjacent taps; i.e. the slider voltage varying with time ' looks like ' a curve of temperature versus depth.

Vol 6 1959

In fact, this voltage may be impressed on one axis of an oscilloscope, the other axis of which is driven synchronously with the interpolating potentiometer, and a BT type trace occurs. Fig. 9 shows three such traces as insets.

The interpolating potentiometer is directly coupled to the shaft of a helical facsimile recorder and is driven by the helix drive motor at a steady (but selectable) rate. This recorder utilizes the rotation of a helix, mounted on a drum, to produce a uniform motion of a ' contact point ' between the helix and a straight edged electrode across the emerging chart paper. A voltage between the helix and this ' blade ' darkens the paper electrochemically.

The position of the contact point is then equivalent to the angle of rotation of the potentiometer, which is equivalent to depth, and it remains only to sample the potentiometer slider voltage and mark the record when this voltage corresponds to an integral temperature (or tenths or five hundredths thereof). This is accomplished by balancing the slider voltage against the voltage from another potentiometer which is driven to balance by a servo amplifier/motor combination (similar to a standard self-balancing potentiometer). This in effect converts the scanning potentiometer voltage (which is proportional to temperature) to a shaft rotation of the servo system. By proper design of the servo gearing and adjustment of the balance voltage it is possible to make one shaft rotation of the servomotor exactly equal to one degree centigrade. A disc is then attached to the servomotor shaft which has slots corresponding to the rotation equivalent to 1°C, 0·1°C and 0·05°C (Fig. 4). Light passing through these slots illuminates one of three phototransistors which provide pulse signals when isotherms are passed. These are amplified and used to mark the recorder.

The servomotor which drives the balance potentiometer is of special construction having very low inertia, with the result that it can attain full speed from a stopped condition in 0·05 secs. The associated servo amplifier is also of special design, as it must have a balanced differential input and present a very high impedance to the scanning potentiometer.

The balance potentiometer is a three-turn, high resolution, high linearity unit. It is energized by the same voltage source that polarizes the thermistor circuits (Fig. 5). Therefore, voltage drifts do not affect the accuracy of the system. The slight nonlinearity of the voltage divider output as a function of temperature is corrected in the balance circuit by adding four taps to the balancing potentiometer. By impressing certain voltages on these taps the voltage distribution along the winding of the balance potentiometer is forced to agree more closely to Fig. 6 than does a linear voltage distribution. While this correction is small it permits the servomotor shaft angle to follow the temperature distribution with an error of less than 0·02°C.

The servo balance system must ' snap back ' from the voltage equivalent to the temperature at the bottom of the chain to that of the surface, at the end of each sweep. Time for this 'snap back' is provided by using only 270° of the interpolating potentiometer winding for the thermistor taps and having the recorder helix made to record the full width of the paper with 270° of drum rotation.

The speed at which the servomotor must turn to follow a particular temperature structure is determined by the steepness of the structure and the selected speed of the recorder drum, the scan rate. There is, however, a maximum speed at which a servomotor can turn, and this determines the maximum allowable scan rate for a given temperature structure. A synchronous rectifier is provided that operates on the servomotor control voltage. This yields a d.c. voltage whose amplitude is proportional to motor speed and whose polarity is determined by direction of rotation. This voltage is used to light a neon bulb if the motor is near full speed, indicating that the scan speed must be reduced.

The same voltage is used to tint the recorder paper any time the servo system is moving in the ' hotter ' direction. Thus, a band appears on the chart that indicates the presence of the temperature inversion in the water column. With this feature, it is only necessary to know the identity of one isotherm appearing on the record. Both the scanning potentiometer and the servo balance system drive ' slave ' potentiometers whose outputs are put on meters. Thus, the scan can be stopped at the position on the paper corresponding to any line (isotherm) and the meters turned on to identify that isotherm and its depth.

(5) EXAMPLES OF DATA

The equipment described above (or some of its prototypes whose operation was essentially identical with it but lacked automatic inversion detection) has been towed extensively over the last two years. Many trips have been made along the East Coast of the United States and in the Caribbean

Sea. An Atlantic crossing from Newport, Rhode Island to Gibraltar via the Azores was made in July of 1958, and the equipment was used in the Mediterranean Sea during the summer of that year. The return crossing from Belfast, Ireland to Argentia, Newfoundland and thence to Boston, Masasachusetts was made in September, 1958. Following this, several other trips have been made in the Western Atlantic including a New England to Bermuda passage and return in December, 1958.

Fig. 7 shows a typical record for the slope water south of New England in the fall. Such a record is also typical of the Sargasso Sea (with, of course, different temperatures) during the months when a pronounced seasonal thermocline exists there. It may be noted that the isotherms are quite level and uncomplicated, which is in sharp contrast to the picture shown in Fig. 8. This recording was made about fifty miles east of Lisbon, Portugal, but is typical of most of the eastern North Atlantic during the summer and early fall. In this figure there are pronounced indications of periodicity which may be internal waves. However, without further data it is impossible to tell whether these are actually moving waves or stationary temperature structure. Fig. 9 shows a recording taken near the centre of the Western basin of the Mediterranean Sea, which is typical of the complexity encountered there. The three insets on the record are photographs of the BT-like trace obtained on an oscilloscope as mentioned above. It may be noted that a pronounced deepening of the isothermal layer and sharpening of the thermocline occurs at about three miles, and this is followed at five miles by a shallowing and further sharpening of the thermocline. Fig. 10 is a short section showing a passage out of the Gulf Stream into slope water east of Jacksonville, Florida. The slope of the warm water overlying cooler coastal water is apparent. The series of inversions which occur throughout this record at about 175 ft are caused by a thermistor which is slightly off calibration being towed at that depth. Fig. 11 is a similar crossing of the edge of the Gulf Stream south of New England in December. In this case the slope water to the north was isothermal to about 350 ft. The complex thermal structure which occurs just before leaving the stream presents a problem in isotherm identification and contouring which the prototype instrument could not handle.

Acknowledgements—The authors wish to acknowledge their indebtedness to their colleagues at the Woods Hole Oceanographic Institution who have assisted in the design, construction and use of this equipment. In particular, we are indebted to Mr. Charles H. Wilkins, who sorted and calibrated the thermistors and assisted in construction of the equipment. The work of Mr. Charles P. Cook in preparation of the figures is gratefully acknowledged. The patience and assistance of the officers and men of the U.S. Coast Guard Cutter *Yamacraw* and the USNS *Chain* are greatly appreciated.

Woods Hole Oceanographic Institution
Woods Hole, Massachusetts.

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Vol
6
1959

LETTERS TO THE EDITORS

Proposed International Indian Ocean Oceanographic Expedition, 1962-1963

(*Received* 4 *January* 1960)

ONCE again, oceanography is approaching a crossroads. Two recent events strongly indicate that the turning point has nearly been reached : (1) the *International Geophysical Year*, 1957–1959, when about 40 nations and nearly 60 research vessels of all sizes took part in systematic surveys in the world oceans, and (2) the *International Oceanographic Congress*, 30 August to 11 September 1959, in the United Nations building under the sponsorship of the American Association for the Advancement of Science, U.N.E.S.C.O. and the Special Committee on Oceanic Research of the International Council of Scientific Unions. The latter was ' to provide a forum for the exchange of information and ideas ' on the most recent developments in the study of the oceans in all its aspects, physical, chemical, geological, geophysical, biological, etc. The sessions of the Congress did more than this, because of the excellent public relations throughout. It consequently provided the necessary impetus to arouse great public interest in oceanography and subsequently various national and international organizations have shown a greater willingness to sponsor and finance further research on an expanded scale.

Within the span of fifty years oceanography has grown from a branch of geography, which was chiefly descriptive and statistical, to a fully-fledged discipline. In the early days (1910-1914) at the University of Berlin, we learned to handle water bottles, reversing thermometers and current meters in a lake only 37 m deep near Potsdam! About the same time, Bergen became the Mecca for oceanographers to study the physical-dynamic methods of FRIDTJOF NANSEN, BJØRN HELLAND–HANSEN, V. BJERKNES and J. W. SANDSTRÖM. With the small research vessel, *Michael Sars* meticulous systematic measurements were made on many sections and stations in the Norwegian Sea. From dynamic analyses of these data, a new era was forecast in the classic ' The Norwegian Sea ' by NANSEN and HELLAND–HANSEN. In contrast to this modern approach in a limited area, oceanographic voyages in the open oceans from the time of the *Challenger* Expedition (1872–1876) until 1925 had more the character of isolated, random tests. Only by a combination of the results from several such expeditions could one form the first rough picture of the bathymetry and the stratification of the water masses in the deep oceans. Such work was purely descriptive, qualitative and without a clear understanding of the circulation or mixing processes. This first period then may be termed the *Era of Exploration*.

A new epoch in the study of the oceans was initiated in 1925–1927 on the *Meteor* Expedition with the 14 cross-sections of the Atlantic Ocean from 20 °N to the Antarctic ice edge. By this series of closely spaced stations with subsurface observations at standard intervals down to the greatest depths, it was possible to accumulate data in a systematic way on the bathymetry, on the bottom sediments, and on the distribution of the chemicals and plankton as well as the stratification of the water masses. For the first time it became possible to obtain a reasonable quantitative approximation for the circulation and for certain mixing problems. Since 1930, the *Discovery II* (2100 tons) has obtained additional systematic and fundamental hydrographic data in the Antarctic like that of the *Meteor* (1178 tons) and since 1931, the *Atlantis* (460 tons) has undertaken similar work chiefly in the northwest Atlantic. This second is the *Era of Systematic and Dynamic Ocean Surveys*. To what extent was it possible to proceed from qualitative results to more quantitative ones in the central problem : the *Deep-Sea circulation* ? In reviewing the last dynamic report of the *Meteor* Expedition (*Trans. Amer. Geophys. Union*, Vol. 39, No. 6 pp. 1171–1172), HENRY STOMMEL stated that this analysis of the *Meteor* data ' has been able to bring his [WÜST's] whole concept of the circulation of the Atlantic into sharper focus, and that there can hardly be any doubt that he has been able to delineate the main features of this great natural phenomenon.' One of the more valuable contribu-

tions to this analysis was the first fundamental study of the topography of the *reference layer* carried out by Prof. Dr. A. DEFANT. In the mentioned reveiw STOMMEL continues with the following words 'Since the time of the *Meteor* Expedition, much oceanographic survey work has been done but never in quite the same systematic way, nor have the results on any of the later expeditions been written up in such a thorough fashion; indeed the *Meteor* work was the model for the recent I.G.Y. surveys of the Atlantic' (STOMMEL, 1958, p. 1171).

During the I.G.Y. (1957–1959) through international cooperation in the Atlantic of the research vessels, *Crawford* (U.S.A.), *Atlantis* (U.S.A.), *Discovery II* (U.K.) and Argentinian survey ships, cross-sections were made for every 8° of latitude between 48 °N and 48 °S. This Atlantic I.G.Y.–Programme was in part a repetition of the *Meteor* Expedition and in part an extension of it. There were, however, some important variations in planning the grid of stations. The *Meteor* profiles, on the assumption that deep geostrophic currents were parallel to the isobaths, were at right angles to the continental slopes and the slopes of the Mid-Atlantic Ridge. Thus, particularly between 5 °S and 20 °N these sections ran in a south-west north-east direction. Since the earlier assumption cannot be proved in all details and for more adequate comparison of the profiles, the I.G.Y. profiles were run east-west (or vice versa) along the degrees of latitude. The number of stations was increased to reduce the distance between them. Furthermore, instruments were spaced at 200 m intervals in sampling at depths greater than 2000 m, rather than at the standard 500 m intervals. In addition, in the northeast Atlantic, the so-called *Polarfront–Programme* extended the area even farther north through the cooperation of nine nations by traversing a more diagonal network at two different seasons.

According to FUGLISTER, preliminary results indicate that 'Over most of the deep water in the tropics the recent *Crawford* sections show virtually no difference from those of the *Meteor* some 30 years ago, thus verifying in a most convincing manner, the assumption by WÜST that the deep circulation is in a climatologically mean steady state' (STOMMEL, 1958, p. 1172). It also appears that the *Crawford* sections will confirm 'the amazing concentration of deep currents in limited, narrow, high velocity western boundary currents often lying right on the continental slope of South America,' as calculated for the *Meteor* data (WÜST, 1957). In the eastern central Atlantic, the small calculated velocities of the meridional components (above and below the arbitrary *reference level*) cannot all be permanent ocean features because there may be momentary disturbances in the density field caused, for example, by winds and internal waves. But 'when transports per unit depth are computed for the full width of an ocean basin, or for the whole western or eastern trough, the contribution of these ambiguities is minimized, as WÜST shows, and one is bound to agree with the author that these integrated transports compare most gratifyingly with the expectations of water mass analysis, each computed current maximum (the intermediate water excepted) coinciding nicely with the centre of a particular water type' (STOMMEL, 1958, p. 1171).

As a consequence, despite some reservations on theoretical grounds, calculations of the geostrophic currents can be very useful in providing a reasonable first approximation of the climatological mean steady state of the deep circulation. Later this will have to be verified by systematic direct measurements using a device such as SWALLOW's neutral buoyancy float. The first successful experiments have indeed already been made in the northwest Atlantic (SWALLOW and WORTHINGTON, 1957).

As a result of the accomplishments during the I.G.Y. programme an excellent systematic survey has been completed for the entire Atlantic. The first practical step is the observational work at sea, with the pleasures of exploration and visits to strange ports. The second step which will provide the new results anticipated for deep circulation is more tedious. It is the duty, however, of oceanographers to complete the analysis of the tremendous amount of new data that has been accumulated. This task should also 'set the highest standards for publication of the results of oceanographic investigation, complete documentation of the actual original data obtained, the methods of observation, definitive and handsome atlases and complete and explanative studies and interpretation of the meaning of the results' (STOMMEL, 1958, p. 1171). Such tasks occupy many years and require much patience as well as funds for assistants and the expense of publication. It is important that some estimate be made of requirements such as these and that they be obtained in advance. There are all too many examples of expeditions with very few reports of the accomplishments, because such expenses were not arranged for in advance.

During the New York Congress, there were meetings of the Special Committee on Oceanic Research to discuss plans for future deep-sea expeditions, in particular, the *International Indian*

Vol 6 1959

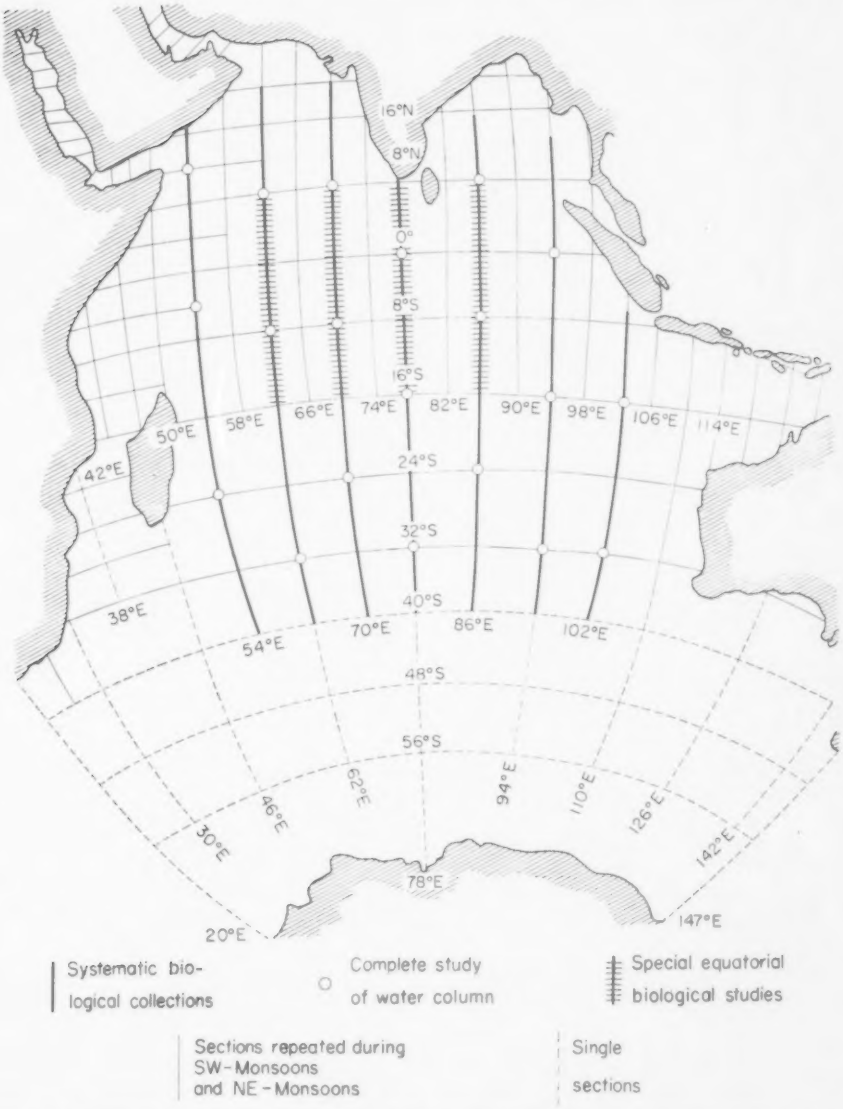

FIG. 1. Proposal of schematic sections for a first systematic oceanographic survey of the Indian Ocean. It is based on an original chart prepared by the author (see description in text). It has, however, been modified to conform with the most recent plans for the collection of biological data (prepared by Dr. N. B. MARSHALL). This revision is dated 21 January and is attached to the preliminary prospectus of the International Indian Ocean Expedition, which has been widely disseminated.

Ocean Expedition, 1962–1963. Other than to stress the importance of sufficient funds for the completion of a survey as I have just done in the previous paragraph, I have considered a possible plan from a purely scientific point of view, disregarding all such practical problems of organization as the number of ships required, the cooperation of interested nations, etc. I have sketched an idealized schematic network of the sections and stations (Fig. 1) for a first systematic oceanographic survey of the Indian Ocean between 30 °N and the Antarctic ice limits including the Red Sea and Persian Gulf. In order to provide a good comparison, it is suggested that the profiles be run along the same latitudes (0°, 8°, 16°, 24°, 32°, 40°, 48° and 56°) as in the Atlantic during the I.G.Y. In the open ocean the distances between stations may be 50 to 100 miles apart, but within the stronger current systems (Somali Current, Equatorial currents, etc.) the interval should be reduced to only about 20–40 miles. The observations should be taken at the same depths as during the I.G.Y. in the Atlantic (as described above). The classical measurements must be supplemented whenever possible by modern instruments including direct current measurements in the surface layers (perhaps with the G.E.K.) and at greater depths (perhaps with SWALLOW's neutral buoyancy float). As I have already explained at some length above, results of the *Meteor* Expedition 1925–1927 and the Atlantic I.G.Y. programme 1957–1959 indicate that a similar programme in the Indian Ocean should provide mean approximations for determining the stratification and circulation problems. In addition, such a systematic survey would provide the much-needed information on the physical, chemical, biological and marine meteorological conditions. Although it was possible to make one survey to cover most of the Atlantic, as described above, in the Indian Ocean due to complete reversal of the wind and current systems, the stations will have to be repeated during each monsoon (i.e., in January through March and in June through August) between 24 °S and 32 °N. Because of the dominating zonal components in the warm waters of the Indian Ocean, this should be done not only from east to west, but also from north to south for each 8° of latitude. Furthermore, the interesting Somali Current system with its very high velocities, its reversal in direction and shifting of its discontinuities (divergences and convergences) from one monsoon to the other require a further narrowing of the distance between cross sections in that area to 4° of latitude. It is recommended that this grid be used in all coastal areas. At least some of the sections should be repeated during the winter in the stormy regions of the Indian Ocean and in the Antarctic Basin.

To some, this proposed network of stations for the first systematic survey of the Indian Ocean may appear to be too schematic and too utopian; to others, it will seem to be obsolete because it will not yield sufficiently synoptic data. It will, of course, have to be modified depending on the number of vessels and on the time available. Furthermore, in proposing this survey, the second period in oceanography (i.e., the *Era of Systematic and Dynamic Surveys*) is in effect being prolonged. During the Oceanographic Congress, I often heard the phrases ' only descriptive oceanography,' ' more or less routine work,' etc. used in a rather deprecating manner, particularly by theoreticians. In part they are right; surveys of this sort using the older classical methods will never describe true dynamic features. However, in the present state of our knowledge – or better ignorance – of the Indian Ocean, we should be contented with a survey of the caliber of that in the Atlantic (with the modifications described above). It is said that younger oceanographers avoid routine work and follow their own inclination and interests. This is indeed important for developing new ideas, new methods and better instrumentation for use in future deep-sea expeditions – an aspect of oceanography in which the American and British oceanographers now excel.

In my old-fashioned view, I believe that it is desirable to begin the Indian Ocean survey (and sometime also one in the Southern Pacific Ocean) with a systematic survey similar to that already completed in the Atlantic during the I.G.Y. For this reason, I have indulged in a considerable amount of reminiscence to illustrate why such a survey seems so important to me. When the stratification phenomena, circulation processes and the mean steady state are known for the world oceans I will concede that we are truly entering the *Era of synoptic oceanography* (or perhaps better quasi-synoptic). We will be ready for synoptic surveys of current systems with many research vessels making repeated synoptic surveys of some particular area. Just as the second *Era* was foreshadowed by the group working in Bergen about 1910 the surprising new results on the dynamic features of a convectional current resulting from Operation *Cabot* in the Gulf Stream system reveals what we may anticipate in this third *Era*.

In closing, I would like to quote HENRY STOMMEL, once again, because of his great insight into the theoretical problems of dynamic oceanography. On convening the morning lectures on the ' Deep-

Vol 6 1959

Sea ' in the General Assembly Room of the United Nations on 4 September 1959, he said ' There is, however, one aspect of knowledge about the ocean which has been developed during the last forty years and in which some of the best work is still being done : the *systematic surveying and mapping of certain scalar quantities in the ocean* : temperature, salinity, and density of the water. This difficult, painstaking and sometimes tedious work is *the very foundation upon which all other study of the ocean must be built*. There are still areas of the world ocean, where the available data are incomplete.' I would like to add that this is especially the case in the Indian and South Pacific Oceans.

Acknowledgement—The author wishes to thank the U.S. Editor, Dr. MARY SEARS, for her help in making the English text more idiomatic.

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Vol. 6 959-60

Indian Ocean vertical temperature sections*

(*Received* 19 *February* 1960)

THE SCOR Committee for the planning of the multi-ship international oceanographic survey of the Indian Ocean in 1961-62 requested information from the Bathythermograph Section of Scripps Institution of Oceanography concerning the distribution of bathythermograph temperature observations in the Indian Ocean.

The following information and vertical temperature charts were distributed to Committee members in New York in September, 1959. Since these data are unavailable to the large number of scientists who may take part in this expedition, it is hoped that Deep-Sea Research will bring them to the attention of interested persons.

The Scripps Institution of Oceanography, University of California, La Jolla, and the U.S. Navy Hydrographic Office maintain a file of bathythermograph temperature observations in the Pacific and Indian oceans. These observations have been taken between 1942-59 by U.S. and British Navy ships, by U.S. and British research vessels and by a few merchant ships. Figure 1 shows the distribution of bathythermograph observations in the Indian Ocean.

The majority of the observations are in the northern parts of the Indian Ocean along the sea lanes between Singapore, Ceylon and the Red Sea. Another large group fall on the route between Ceylon and the Persian Gulf. There are few latitudinal ship tracks and few longitudinal tracks in the vicinity of the Equator and to the south of it.

From the available data, eleven cross-sections of temperature have been drawn, five of which are latitudinal sections. Figure 2 is a location chart of the sections. Insofar as possible, the sections were selected to show differences between monsoon seasons and between years. The only latitudinal bathythermograph sections which were omitted are those made by the *Vema* of the Lamont Geological Observatory, Columbia University, and the *Atlantis* of the Woods Hole Oceanographic Institution in 1958.

Figures 3, 4 and 5 are latitudinal sections. Figures 6, 7 and 8 are longitudinal sections. Since none of the sections was exactly East-West or North-South, both latitude and longitude scales are

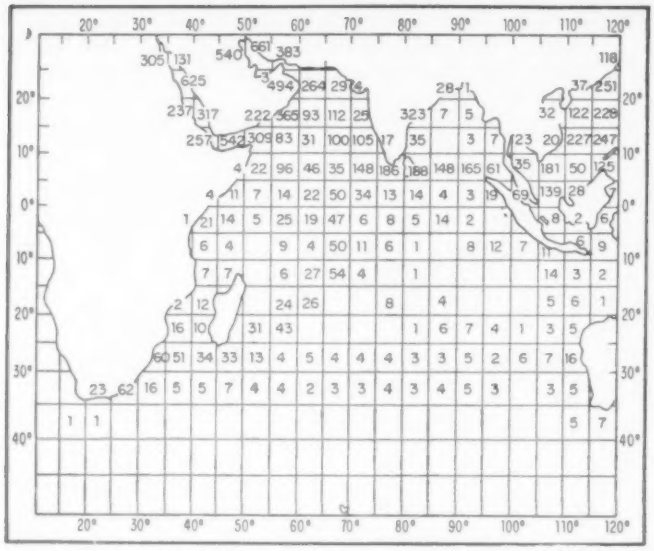

FIG. 1. The total number of bathythermograph observations taken in each 5-degree quadrangle of the Indian Ocean (5,625), Red Sea (1,872), Persian Gulf (1,587) and So. China Sea (1,901), as of January 1, 1960. These include data processed at Scripps Institution of Oceanography, Woods Hole Oceanographic Institution and the U.S. Navy Hydrographic Office in the years 1942-1959.

Vol 6 1959

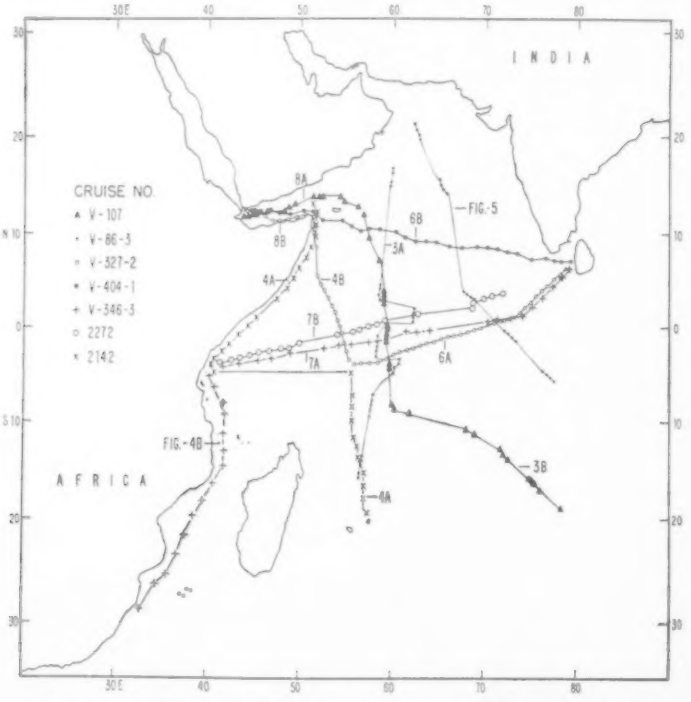

FIG. 2.	V–107–1	VI, VII–44	S.S. *Francis Lewis*
	V–86–3	I, II, III–45	U.S.S. *Fort Niagara*
	V–327–2	IV, VII–52	H.M.S. *Challenger*
	V–404–1	X, XI–53	U.S.S. *Caperton*
	V–346–3	V, VII–54	U.S.S. *Soley*
	2272	XII–56	H.M.S. *Cheviot*
	2142	V, VI, VII, IX, X, XI-57	H.M.S. *Ceylon*

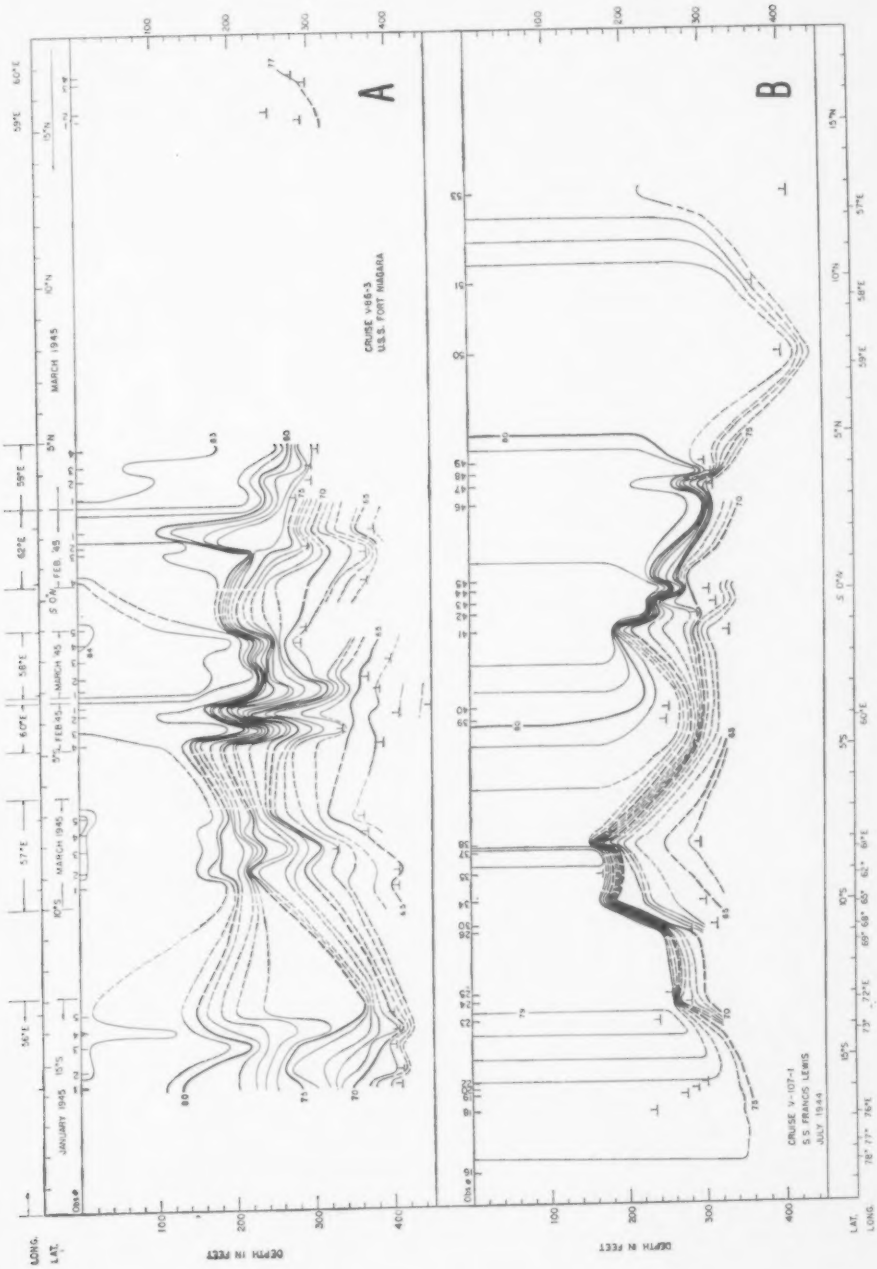

FIG. 3. Latitudinal Sections.

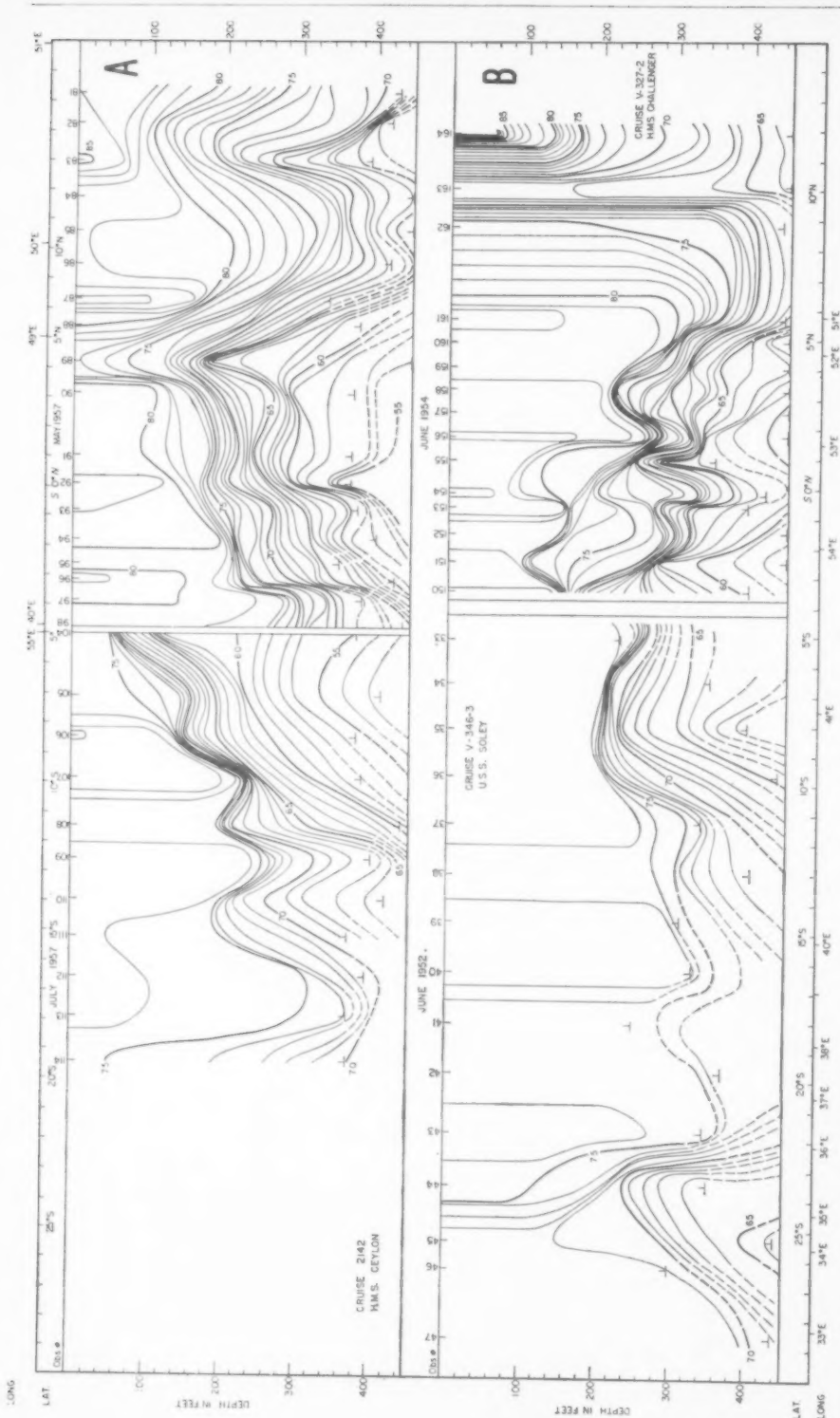

FIG. 4. Latitudinal Sections.

Vol 6 1959

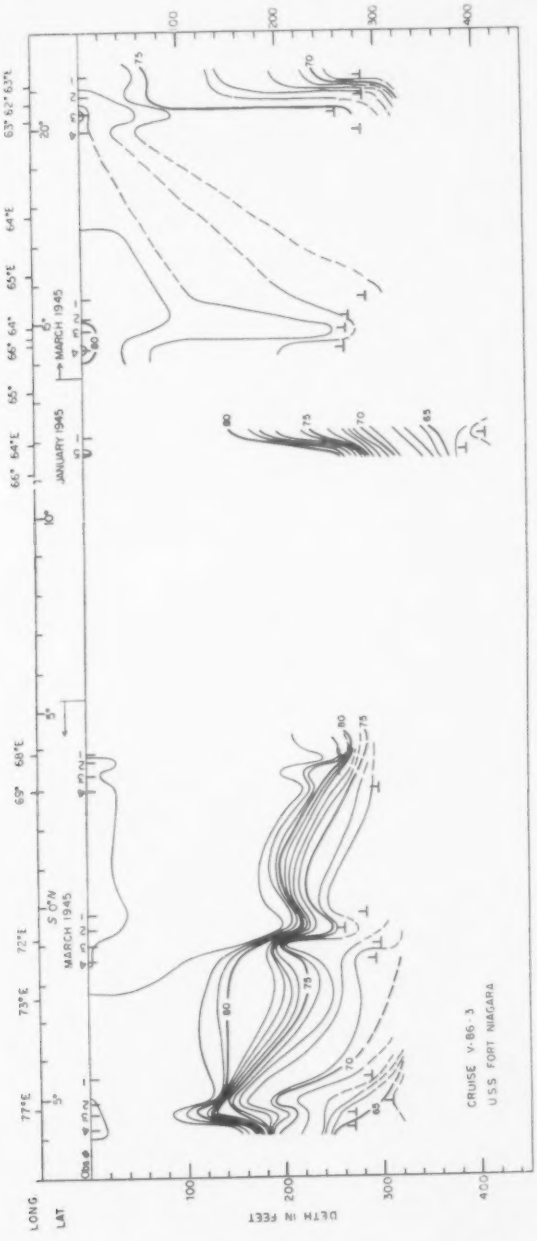

FIG. 5. Latitudinal Section.

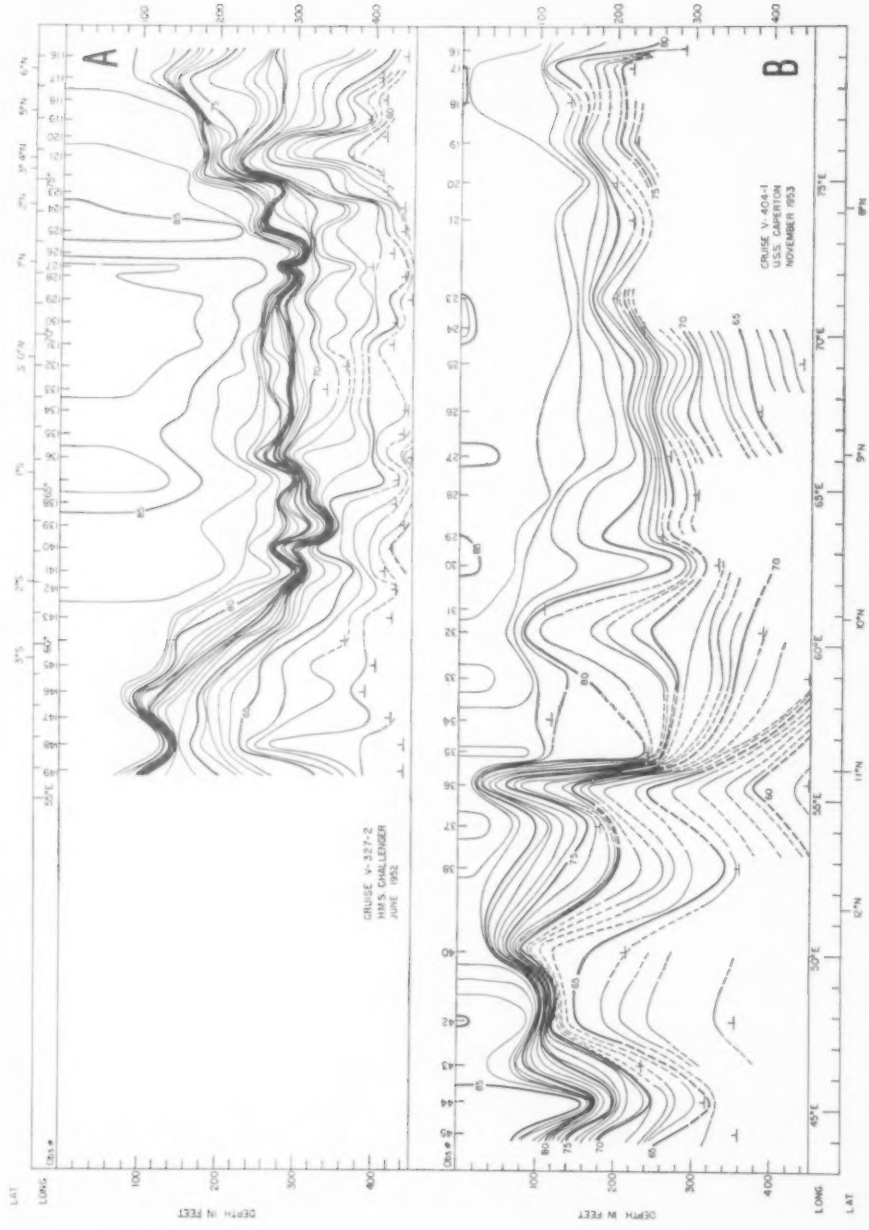

FIG. 6. Longitudinal Sections.

Vol 6 1959

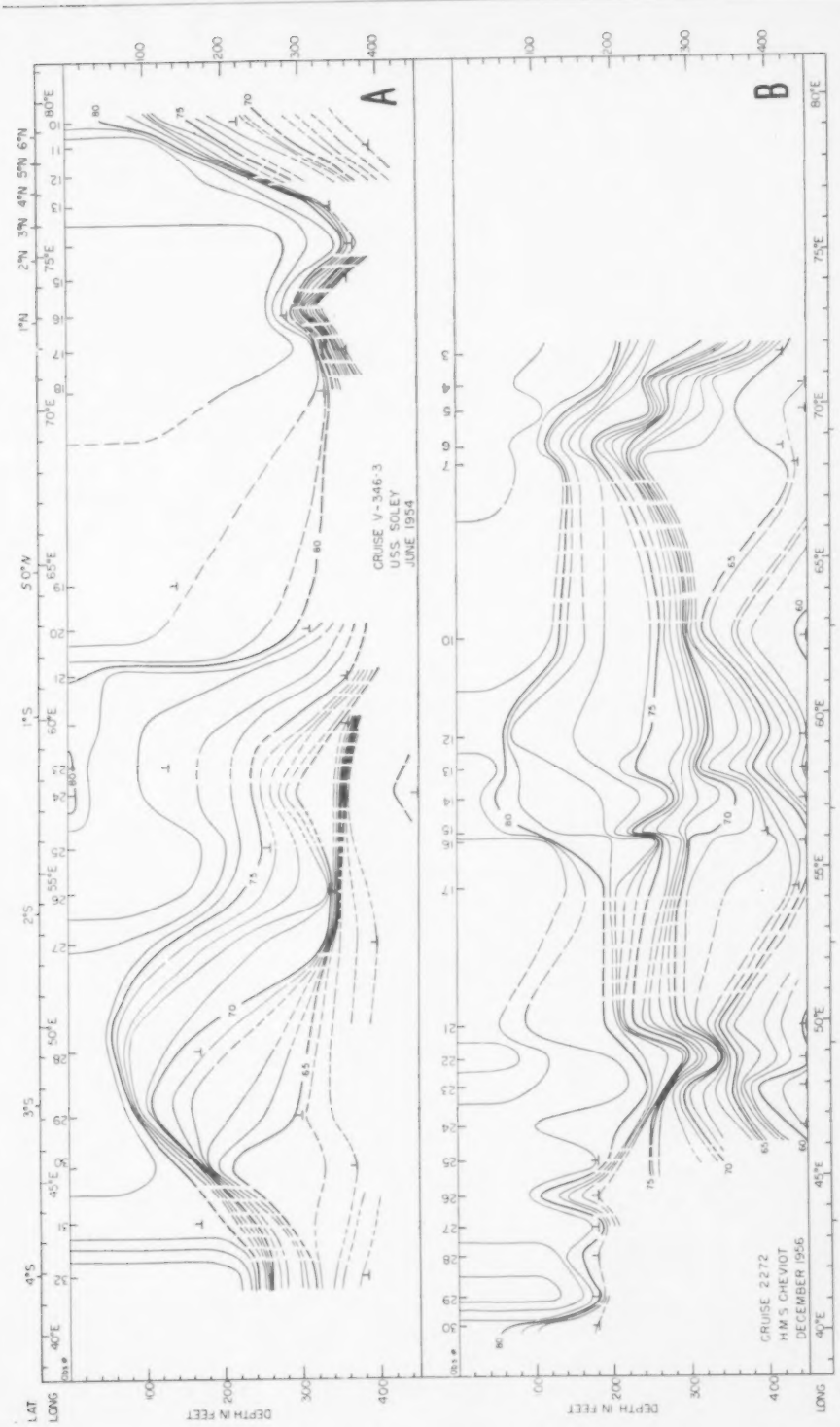

Fig. 7. Longitudinal Sections.

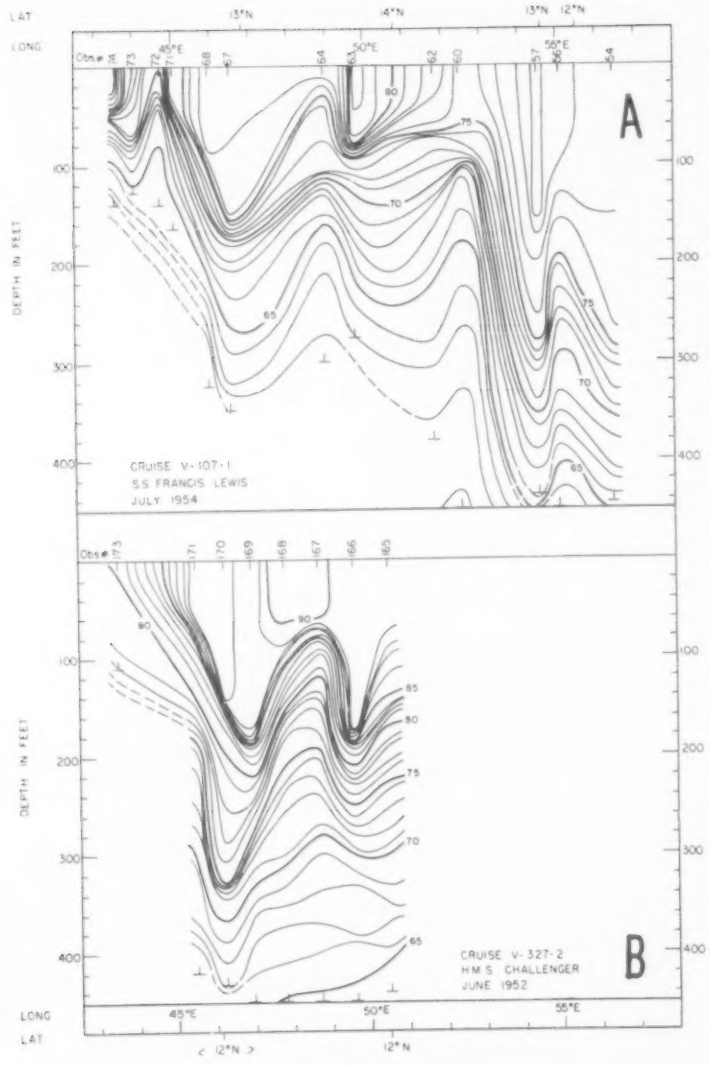

Fig. 8. Longitudinal Sections.

Vol 6 1959

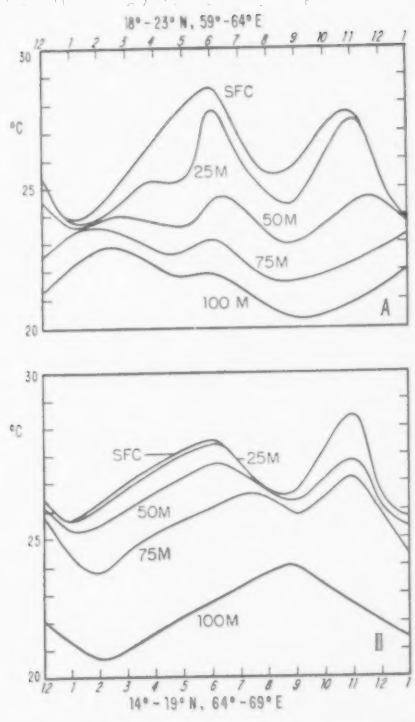

FIG. 9. Annual variation in the Arabian Sea from bathythermograph observations.

indicated on the margins. The inner evenly-spaced scale is the one against which the observations are plotted. The outer irregular scale locates the section with respect to the second coordinate. The inverted 'T' indicates the depth of the observations whose geographical location is indicated at the top of each section by a tick and the observation number. Dashed lines have been used when isotherms are connected below the depth of the observations, or where distance or time between observations is large. Sections have been broken when distance or time between observations is excessive or when the observations were taken by different ships. All sections have been reproduced on the same scale.

Figures 9a and 9b are charts showing the annual variation in the Arabian Sea. These charts are based on bathythermograph data analyzed by JUNE G. PATTULLO for use in her study of *Seasonal Oscillation in Sea Level* (PATTULLO, *et al.*, 1955). They are reproduced here with her permission. The data have been further smoothed to reduce the effect of erratic distribution in time and space.

Acknowledgements—This paper represents results of research carried out by the University of California, under contract with the Office of Naval Research. Reproduction in whole or in part is permitted for any purpose of the U.S. Government. This research also received support from the California Cooperative Oceanic Fisheries Investigations, a project sponsored by the Marine Research Committee of the State of California.

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**Contribution from the Scripps Institution of Oceanography, New Series Number.*

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Vol
6
1959

BOOK REVIEWS

The Floors of the Oceans—I. The North Atlantic, B. C. HEEZEN, M. THARP and M. EWING. Spec. Pap. Geol. Soc. Amer. **65** (1959).

Two events during this year have helped to raise the status of oceanography from a little known field of study to an important branch of science on a footing with nuclear research and outer space exploration. The United States have pressed for a three-fold expansion in their oceanographic programme and have already done much to initiate it. Secondly the International Oceanographic Congress at New York, the first of its kind, has brought together under a common title, an amazingly wide collection of papers varying from the circulation of oceanic waters, to the ecology of a little known fish in the waters of a little-known bay. This wide range of interests, both topical and regional, have one thing in common – the sea.

Just as on land, the studies of botany, zoology, geology and many other branches of the natural sciences are intimately related to a knowledge of environment and in particular to the regional geography, so in the sea these studies cannot be divorced from a knowledge of the nature of the sea floor. This boundary is both the upper limit to the earth's crust and the lower limit to the ocean and so has a doubly important place in oceanography.

Spasmodic research into the nature and shape of the sea floor has gone on for nearly a century but it was not until the last decade that enough work had been done on an ocean-wide scale to enable a treatise such as this by Dr. HEEZEN to be written. The Lamont Geological Observatory has been foremost in the field of sea floor exploration in the Atlantic Ocean and the team working under Professor MAURICE EWING have used many different techniques in attacking the problem. Dr. HEEZEN's treatise and the physiographic diagram are not the work of one man, but are the collective results of this team and represent an attempt to portray in as simple and direct a manner as possible what is the nature of the Atlantic floor.

The physiographic diagram, which covers the North Atlantic and which is the first of a series of such diagrams covering the world's oceans, portrays the topography of the sea floor in a semi-pictorial way as if viewed obliquely from the south. It has been based on continuous echo sounding profiles together with an intelligent interpretation of sea floor features to fill in areas where no such soundings lines exist. The merit of this type of presentation is that it combines the advantages of profiles, which can show peaks, valleys and slopes too small to appear on contoured charts, with the areal distribution of the profiles, giving a three dimensional effect. It's purpose is to illustrate the nature of the topography rather than to show the depth of water at any given point. It is therefore complementary to, rather than a substitute for, a contoured chart. It may be argued that the vertical exaggeration of 20 : 1 gives a too distorted idea of the slopes and the sharpness of features. However, an oceanographer has constantly to be aware of this danger and mentally to correct for it, since it is common in all presentations of topography. It is a pity, however, that due to the relative position of features that the one ocean trench in the North Atlantic is obscured by the neighbouring islands.

The physiographic diagram forms the background on which the analysis of North Atlantic topography is based. In the Special Paper, Dr. HEEZEN has revised the classification of submarine topography in the light of the regions revealed in the diagram. He has rejected the ' bathymetric ' classification in favour of a ' geomorphic or textural ' one. The three major divisions – continental margin, ocean-basin floor and mid-oceanic ridge – are sub-divided into categories of provinces and sub-provinces. Each of these is described in detail with ample illustrations of profiles, charts, tracings and photographs of echo sounding records. The discussion on each major division concludes with a summary of all geophysical measurements that have been made and an interpretation of the data with regard to its structure and origin. Clearly in an undertaking of this nature, many of the controversial topics have to be dismissed very briefly but a good bibliography leads the reader well into such matters.

There are some important new theories presented in the discussions, perhaps the most significant of these being with regard to the structure of the continental shelf. Evidence, albeit rather tentative, is presented for the correlation of structural benches on the continental slopes with geological strata as found from well logs. Many such benches have been correlated along the slopes off the north-east United States. The implication of this and other evidence is that the continental slope is not a depositional feature but one due to preferential erosion revealing continental stratification and that the base of the continental slope is a structural discontinuity. Confirmation of these conclusions must lie in sampling the rocks on the steepest parts of the slope especially on submarine canyon walls.

The ocean-basin floor is divided into two provinces, the abyssal plain and the abyssal hills, but there does not appear to be any category for the undulating basin floor representing rough basement topography covered by sediment not transported by turbidity currents. The abyssal plains, by definition, cannot have slopes greater than 1 : 1000 and the abyssal hills are characterized by their discrete nature and steep slopes, giving rise to the multiple echo type of record illustrated in Plate 9, Fig. 4. There are in fact many regions in the basins where neither description applies. Such a region lies between the Azores and the Horseshoe seamounts and also east of the Southeast Newfoundland ridge (cf. Fig. 30, Profile VI).

The provinces of the mid-Atlantic ridge region are the most difficult to allocate and it is only with great caution that Dr. HEEZEN has defined the upper, middle and lower steps. It is difficult, indeed, to correlate the scarp zones which form the province boundaries. The median rift valley is demonstrated to run the whole length of the ridge although in places there appears to be more than one valley. Detailed surveys by other workers have shown that at least in one place the valley is closed off at one end and it seems more probable that a series of valleys sometimes connected and sometimes overlapping, run parallel to and near the peak of the ridge.

These criticisms, however, point to the wide coverage of this work and to the number of unsolved problems that remain. The author realises this as well as anyone and states very clearly where our knowledge is lacking and our theories inadequate. This is the challenge of the book. It provides the framework into which the more detailed regional studies can be fitted and shows where future work can most profitably be carried out.

The basic problems of the permanence of ocean basins versus continental drift, of the structural development of the large features of the oceans such as the mid-ocean ridges and of the formation of the continental shelf all require a sound knowledge of the present morphology of the sea floor, since this is the immediately available evidence in which many clues can be found. The next stage is the much more difficult one of high quality sampling of the principal rocks that comprise the oceanic crust, for it is by the analysis of actual samples that the many theories can be proved or disproved. Already work is planned on these lines but because of the inherent difficulties of coping with the three miles of water covering the sea floor, it will inevitably be slow. Nevertheless, with the money and effort that is available in other fields of exploration and with the new techniques that are constantly being developed, there is no reason why many of the problems could not be solved in the next decade.

The book is admirably illustrated with high quality diagrams and plates which will doubtless be used for reference in all future work. Deep-sea photographs illustrate the micro-topography and help in visualizing the quality of the sea floor, and the reproductions of echo-sounding records will help others to recognise typical features. This volume, which is the most important work yet published on the ocean floors, will undoubtedly become a reference book for all branches of oceanography.

A. S. LAUGHTON

J. BOURCART : **Problèmes de Géologie Sous-marine**, Masson et Cie, Paris 1958. 123pp., 980Fr.

ONLY in the last few decades has submarine geology been raised to the status of continental geology thanks to the techniques of sampling, measurement and observations that have been developed. Its importance as a study of the oceanic environment and of the boundary regions of the continent, cannot be over-emphasised, for it fills a large gap in the geological picture of the earth's crust as a whole and provides many clues to the problem of its development.

Vol 6 1959

Professor BOURCART's book does not pretend to be a textbook of submarine geology, but considers some of the problems with which the author and his French colleagues have been concerned, dealing mainly with the continental margin region or " precontinent " as he has called it. After discussing in general terms the problems involved, Chapter I deals with the geological constitution, the seawards limits and the distribution of sediments of the precontinent. The study of submarine canyons, especially in the north-west Mediterranean, leads the author to the conclusion that the precontinent is neither an accumulation of sediments from the continent nor the remains of an eroded continent but the result of marginal flexure that has temporarily lowered the continent below sea level. In the section on the sediments the author's intimate knowledge of the shelf deposits off the French coasts is revealed. The characteristics of coastal and off-shore sands and muds are explained in terms of variations of sea level, and of the currents as well as by the action of the marine organisms and bacteria.

Chapter II deals with the development of shoreline features which can be associated with the Quaternary transgressions and regressions of the sea. These transgressions are the subject of his final chapter. A study of the raised beaches and other features shows that they cannot be correlated with glacial epochs and he suggests a mechanism whereby radioactive generated heat from the earth's interior gives rise to periodic expansions and contractions in the volume and shape of the ocean basins.

Doubtless many of Professor BOURCART's conclusions will not be immediately acceptable to geologists and geophysicists and can only be supported or refuted by further detailed surveys. His generalisations have been made from rather a limited field of data and there appears to be too little distinction between observations made in the Mediterranean and those in the Atlantic concerning crustal movements. It is a pity that the inadequacy of diagrams in the text to illustrate his hypotheses makes it difficult at times to follow his arguments. However, if the book has stimulated further studies and provokes argument, then it has wholly achieved what the author intended.

A. S. LAUGHTON

A. GUILCHER : **Coastal and Submarine Morphology** (Translated by B. W. Sparks and R. H. W. Kneese). Revised English Edition – Methuen 1958, 274 pp., 30s.

THE FIRST three-quarters of Professor GUILCHER's book are devoted to the geomorphology of the narrow marginal strip of the ocean between the greatest depth of wave action and a line a few yards inland. The first chapter gives a summary of the forces engaged in sculpturing the coastline : waves, currents, chemical and biological processes. A short second chapter of ten pages, backed by an extensive bibliography, reviews the difficult study of shoreline changes, particularly those of the coasts of Europe and North Africa, and considers the roles to be assigned in their interpretation to eustatic, tectonic and epeirogenic changes. The remainder of the first part of the book presents a classification of coasts and coastal features and discusses their formation and evolution. With the exception of an excellent condensed account of the coral reef controversy there is little material here of direct concern to students of the deep sea.

The second part of the book (64 pp.) presents an outline of submarine geomorphology. ' The balance between the two parts of the work represents the present state of knowledge.' An introductory chapter mentioning oceanographic research organizations and methods is followed by a chapter on the continental margin and its sediments. Theories of the origin of submarine canyons are reviewed with particular mention of the scouring or erosional effects of turbidity currents and of BOURCART's hypothesis of a widespread post-Quaternary flexure at the continental margin. The final chapter relates to the deep-sea floor below 2000 m and touches on such features as ocean basins, mid-ocean ridges, and guyots, and on the distribution of sediment types. A few paragraphs discuss the nature of island arcs and associated trenches in the light of the work of UMBGROVE and VENING MEINESZ and the possible differences in the distribution of sial and sima in the Atlantic and Pacific Oceans in the light of evidence from surface seismic waves. The more recent American work on these topics is not mentioned.

Professor GUILCHER expresses the 'hope that the submarine section of the book will become rapidly out-of-date and that before long it will be possible to give a fuller, more precise, and less hypothetical, *but still geographical* (reviewer's italics) account of the sea floor.' Already one imagines that a new edition of the book would find a place for a discussion of the interconnecting abyssal plains of the Atlantic and for the more dynamic conception of submarine processes presented by HEEZEN. If, perhaps, the book will make its greatest appeal to undergraduate geographers, oceanographers will find in it an authoritative and non-partisan review of the geographical aspects of the ocean margin. Its value is much enhanced by annotated bibliographies at the end of each chapter which contain references to European (and to a lesser extent, American) literature up to 1954. Some of the forty text-figures have been obscured by overcrowding in the English edition. The clear translation has produced an English textbook.

D. H. MATTHEWS

HENRY STOMMEL : **The Gulf Stream, A Physical and Dynamical Description.** University of California Press, 1958 202 pp. $6.00.

H. STOMMEL, der bekannte Ozeanograph an der Woods Hole Oceanographic Institution hat im vorliegendem Büchlein eine wissenschaftliche Beschreibung der Physik und Dynamik des Golfstromes auf Grund der neuesten Ergebnisse der Beobachtung und Theorie gegeben. Der Golfstrom ist unter den Strömungen der Ozeane die auffallendste, wichtigste und besonders charakteristische. Die Geschichte seiner Erforschung reicht weit zurück, in Zeiten, in denen die Klarlegung der ozeanischen Erscheinungen nur auf einfache Oberflächenbeobachtungen fusste und dadurch für eine Erklärung des Phänomens das Feld für Spekulationen völlig offen war. In neuerer Zeit ist durch den Ausbau der ozeanographischen Messtechnik auf See und durch die geophysikalische Durchdringung der Ozeanographie in allen ihrenn Teilgebeiten die Grundlage zu einer rationellen wissenschaftlichen Erfassung der ozeanischen Erscheinungen im ganzen Meeresraum gewonnen worden. Ich persönlich habe diese neuere Entwicklung der Ozeanographie selbst in den letzten 50 Jahren mitgemacht und kann auf Grund dieser Erfahrungen besonders ermessen, welche Fortschritte in ihr erreicht wurden. H. STOMMEL hat in meisterlicher Weise verstanden, diesen Fortschritt an der Hand des Problems des Golfstromes übersichtlich und klar darzulegen und hat uns damit eine Darstellung desselben in einer Weise gegeben, wofür wir ihm sehr dankbar sein müssen.

Nach einer kurzen historischen Übersicht über die Gedanken, die man sich in früheren Zeiten über den Golfstrom gemacht hat und einer Darlegung der Beobachtungsmethoden, die derzeit bei Golfstromuntersuchungen benützt werden, werden in bündiger Weise die neuesten hydrographischen Beobachtungsergebnisse über den Aufbau und die besonderen Eigenschaften des Golfstromes mitgeteilt und jene Kräfte besprochen, die für seine Ausbildung hauptsächlich in Frage kommen. Die in wissenschaftlicher Hinsicht wichtigsten Abschnitte behandeln zwei Arten der modernen Theorien des Golfstromes. Die lineare Theorie einer stationären strömenden Reibungsschicht steht in Verbindung mit der von W. MUNK ausgebauten Theorie der windbedingten ozeanischen Zirkulation, wofür STOMMEL ja selbst die Grundlagen geliefert hat. Die nicht lineare Theorie von CHARNEY und MORGAN sieht im Golfstrom eine trägheitsbehaftete strömende Grenzschicht und scheint in ihren Ergebnissen den Golfstromeigenschaften am ehesten gerecht zu werden. Die klare STOMMELsche Darstellung wird auf die Entwicklung der ozeaischen Dynamik von nachhaltiger Wirkung sein. Besondere Abschnitte befassen sich mit den auffallenden Erscheinungen der Meanderbildung und der Transportschwankungen innerhalb des Golfstromes und mit ihrer theoretischen Deutung.

In einem besonders interessanten Kapitel entwirft STOMMEL schliesslich die Gedanken einer neuen Theorie der thermo-halinen Zirkulation, die auch die Bewegungen der Wassermassen in den Tiefenschichten der Ozeane in sich schliesst. Der aus dem thermo-halinen Aufbau vorhergesehene Unterstrom unter dem Golfstrom mag damit eine theoretische Erklärung finden. In einem Schlusskapitel wird alles einheitlich zusammengefasst, die klimatologische Bedeutung des Golfstromes hervorgehoben und die Notwendigkeit weiterer Forschungen auf diesem Gebiet betont.

Das Studium des STOMMELschen Golfstrombuches hinterlässt einen tiefen Eindruck über die derzeitige wissenschaftlich gut fundierte Epoche der modernen Ozeanographie.

A. DEFANT

Vol 6 1959

The Cromwell Current

INTRODUCTION

Investigations of equatorial undercurrents

THE PAPERS in this volume were not originally written by invitation; however, the interest shown recently in equatorial undercurrents, as evidenced by these studies, led to the proposal that they be published together. It seemed appropriate to dedicate the number to Townsend Cromwell, one of the discoverers of the Pacific Equatorial Undercurrent,* who was killed in an aeroplane crash on June 2, 1958, while en route to join the *Scot* Expedition at Acapulco.

The investigations reported herein include both the description of recent measurements of a major ocean current and a series of theoretical studies on the nature of this feature. The measurements are outstanding, because they result from one of the few successful attempts to describe the flow of such a current by a profile of direct current observations. The current is remarkable in that, although comparable in transport to the Florida Current, its presence was unsuspected ten years ago; even now, neither the source nor the ultimate fate of its waters has been established. No theory of oceanic circulation predicted its existence, and only now are such theories being modified to account for the important features of its flow.

In September 1951, aboard a U.S. Fish and Wildlife Service research vessel long-line fishing on the equator south of Hawaii, it was noticed that the subsurface gear drifted steadily to the east. The next year Cromwell, in company with Montgomery and Stroup, led an expedition to investigate the vertical distribution of horizontal velocity at the equator. Using floating drogues at the surface and at various depths, they were able to establish the presence, near the equator in the central Pacific, of a strong, narrow eastward current in the lower part of the surface layer and the upper part of the thermocline (CROMWELL *et. al.*, 1954). A few years later the Scripps *Eastropic* Expedition, under Cromwell's leadership, found that the current extended towards the east nearly to the Galapagos Islands but was not present between those islands and the South American continent.

It should be noted that the discovery of the Undercurrent was, in a way, incidental to Cromwell's principal research (MONTGOMERY, R. B. 1959 *Limnol. Oceanogr.* **4**, 228). This was concerned primarily with the physical processes controlling the fertility of the sea. His studies including upwelling at the equator, oceanic fronts, " ridging " in the eastern tropical Pacific, wind stirring in the Gulf of Tehuantepec, the effects of islands and submerged banks, and the nature of the pycnocline as revealed by model experiments. During his brief professional career Cromwell created a fund of ideas, the influence of which is just beginning to be appreciated.

WARREN S. WOOSTER,
La Jolla, 17 December, 1959

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*The Equatorial Undercurrent was named by Cromwell, Montgomery and Stroup (1954). Since Cromwell's death it has often been referred to as the Cromwell Current (KNAUSS and KING 1958 *Nature, Lond.* **182**, 601–602). Although time and the use of oceanographers will eventually decide between these names, the original proposal is more consistent with present practice in naming features of oceanic circulation.

(In press) Pycnoclines created by mixing in an aquarium tank. *J. Mar. Res.*

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Vol
6
1959

Measurements of the Cromwell Current

JOHN A. KNAUSS

(*Received* 15 *August* 1959)

Abstract—Direct current measurements made along the equator in the central and eastern tropical Pacific during the IGY show that the subsurface eastward flow, first described in 1954, is a fast, thin, but major current. As defined by the 25 cm/sec contour, it is 300 km wide and about two-tenths of a kilometre thick; it is symmetrical about the equator. At 140 °W the core of the current is at 100 m. Velocities up to 150 cm/sec (3 knots) were recorded. The average transport based on four sections is 39×10^6 m^3/sec. The current was traced from 140 °W to the Galapagos Islands at 92 °W. It was not found east of the Galapagos at 89 °W. The core velocity was 100 – 150 cm/sec from 140 °W to the Galapagos; the depth of the core shallowed from 100 to 40 m. At the surface and at 500 m the flow was to the west.

Hydrographic measurements suggest that the flow is in geostrophic equilibrium to within half a degree of the equator. The major features of the Pacific Equatorial Undercurrent (Cromwell Current), including its general shape, the depth of the core, which slopes upward to the east, can be accounted for by assuming a geostrophic flow in response to the perturbation in the horizontal pressure gradients caused by mixing through the thermocline at the equator.

Direct measurements of the Cromwell Current show that the current extends from at least 92 °W to 150 °W (3500 miles). Indirect evidence suggests that the current will be found to at least 170 °W (4700 miles long) and perhaps as far as 160 °E (6500 miles long).

INTRODUCTION

Previous Observations

THE Cromwell Current is a major feature of the oceanic circulation whose presence was unsuspected nine years ago, and whose existence was not predicted by any of the theories of oceanic circulation. Apparently the first indication that there was anything unusual about the currents at the equator in the Pacific was in September, 1951, aboard the *Hugh M. Smith*,* when fishing for tuna at the equator south of Hawaii. The long line gear drifted to the east while the surface drift of the ship was to the west. Further evidence for an eastward subsurface current was gathered at five long line fishing stations (*John R. Manning** and *Charles H. Gilbert**) between January and June 1952. In August of that year an expedition under the leadership of the late Townsend Cromwell made twelve direct current measurements in the vicinity of the equator and 150 °W by observing the rate of drift of deep drags with respect to those at the surface. One of these at 2° 54′S showed no eastward current; four between 0° 14′N and 1° 31′N indicated an eastward subsurface current, and one at 0° 14′N showed an eastward drift of more than a knot. Four observations between 1° 50′N and 2° 43′N showed a westward current. In all nine cases, the surface drag drifted to the west. The remaining three observations were made in the equatorial Countercurrent near 7 °N where both surface and deep drags moved to the east. Although there was no way of knowing the exact depth of the drogue, the eastward current was placed in the ' lower part of the surface layer and upper thermocline ' at a depth of approximately 100 m (CROMWELL *et. al.*, 1954; MONTGOMERY and STROUP, to be published).

*Vessels of the Pacific Oceanic Fisheries Investigation (POFI) U.S. Fish and Wildlife Service.

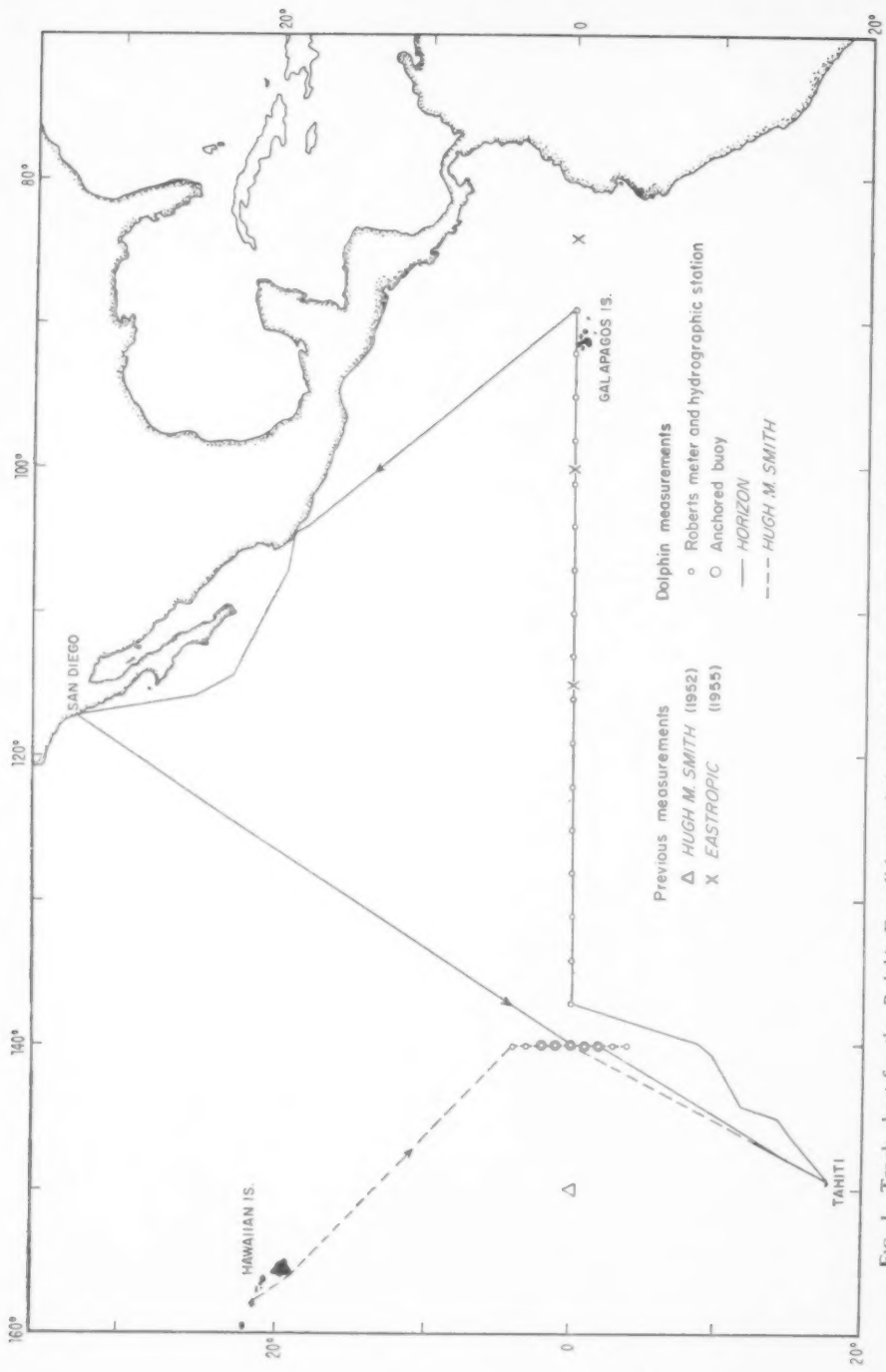

FIG. 1. Track chart for the *Dolphin* Expedition and the location of earlier observations of the Pacific Equatorial Undercurrent.

Vol 6 1959

A further attempt to observe this current with free floating parachute drogues (VOLKMANN *et. al.*, 1956), made by Townsend Cromwell and the author in the fall of 1955 (part of the *Eastropic* Expedition) indicated that the east-flowing current was present at 115 °W and 100 °W, but not at 84 °W.

McGary and Rinkel, with an Ekman current meter between 12 °N and 5 °S along 140 °W (*Hugh M. Smith*, Cruise 33, March 1956) found an eastward subsurface current at the equator (RINKEL, personal communication).

Cromwell suggested the name Pacific Equatorial Undercurrent for this east-flowing subsurface current (CROMWELL *et. al.*, 1954). On June 2, 1958, Townsend Cromwell was killed in an aeroplane crash near Guadalajara, Mexico, on his way to join a Scripps Expedition, just as the joint Scripps-Pacific Oceanic Fisheries Investigation, *Dolphin* Expedition, was completing the study of currents at the equator. It has subsequently been suggested that this east-flowing subsurface current be called the Cromwell Current (KNAUSS and KING, 1958).

Dolphin Expedition

The *Dolphin* Expedition in the spring of 1958 (Fig. 1) was planned to answer the following questions : (1) What is the vertical distribution of horizontal current velocity at the equator ? (2) What is the latitudinal variation of this distribution ? (3) How far to the east can the Cromwell Current be traced ? (4) How does the velocity profile of the current change eastward along the equator ? (5) Is there any marked day-to-day change in the current ? (6) How is the Cromwell Current related to the observed distribution of such properties as temperature, salinity, oxygen or phosphate at the equator ?

For this expedition a new technique for measuring currents in the open ocean was developed. A taut wire buoy similar to that described by ISAACS *et. al.* (1957) was anchored to serve as a fixed reference point. A propeller-type current meter equipped with a depth indicator, a modified Roberts Meter (ROBERTS, 1952), was lowered from the ship at different depths to observe the ' apparent current,' i.e., the water flowing by the current meter. The drift of the ship with respect to the anchored buoy was measured by radar. The true current at any depth was then found by vector subtraction of the ship drift from the apparent current.

For some of the observations a free-floating parachute drogue with the parachute suspended at 800–1000 m was used as the reference point. Although such a drogue is in no sense a fixed reference as is an anchored buoy, the drift of these deep parachute drogues is small compared to the currents measured on *Dolphin*.

*Horizon** proceeded directly to the equator at 140 °W to anchor a buoy, where current observations were made twice a day, either by *Horizon* or by *Hugh M. Smith* during the first half of the expedition. At the same time *Hugh M. Smith* made current measurements with reference to deep parachute drogues at stations along 140 °W at 4 °N, 3 °N, 2 °N, 1 °N, 1 °S, 2 °S, 3 °S and 4 °S to determine the approximate boundaries of the Cromwell Current. On completion of this exploratory section, *Hugh M. Smith* took the measurements at the equatorial station and *Horizon* anchored four more buoys at 2 °N, 1 °N, 1 °S and 2 °S along 140 °W. The positions were chosen on the basis of *Hugh M. Smith* measurements which indicated the Cromwell Current was either very weak or missing at all the stations other than at 1 °N and 1 °S.

*Vessel of the Scripps Institution of Oceanography, University of California.

The equatorial station was monitored continuously by one or the other ship from 6–27 April; the second ship took current observations at the other anchored buoys. Both ships departed for Tahiti, 28 April.

Hugh M. Smith turned to other work after visiting Tahiti, but made a final hydrographic station at the equator, 140 °W on 15 June. *Horizon* returned to the equator and made measurements along the equator between 137 °W and 89 °W at stations 180 miles apart. A 1200 m hydrographic cast and a Roberts current measurement to 300 m were made at each station.

INSTRUMENTATION, TECHNIQUE AND ESTIMATE OF ERROR

Anchored Buoys

Taut wire anchored buoys have been under development at Scripps since 1952 (ISAACS *et. al.*, 1957). Starting at the bottom, the buoy system used on *Dolphin* consisted of a 500 lb anchor, a mooring wire about 4000 m long (0·123 inch diameter, weight 28·8 lbs/1000 ft, 2300 lb breaking strength) attached to a 34 inch steel sphere (positive buoyancy 537 lbs). The length of mooring wire was predetermined to keep the underwater float about 200 m below the surface. Three hundred thirty-five metres of $\frac{3}{16}$ in. wire rope ran from the underwater float to the surface float (an eight foot fibreglass skiff with a light and radar antenna).

By buoying the long length of mooring wire well below the surface, several advantages are gained which help to minimize the extent to which the buoy can swing on its anchor. The first advantage is that the underwater float exerts a buoyant force upon the mooring wire; and if this force can be kept large compared to the horizontal drag forces on the system, there is little horizontal movement of the underwater float. The second advantage of the underwater float over a conventional mooring is that the underwater float is not as subject to forces such as those generated by a heavy sea, and therefore, a light-weight, small-diameter (and consequently low-drag) mooring wire can be used. The surface buoy can also have low-drag characteristics, since it need only be large enough to support the heavier wire used in running from the underwater float to the surface buoy. Finally, the ratio of wire length to water depth of a taut wire mooring can be less than that used with safety in a conventional mooring system.

Roberts meter measurements

A modified Roberts current meter (ROBERTS, 1952) used for these measurements was suspended in a steel bridle with an electric swivel by a single conductor, insulated, steel wire. Twenty-four volts, alternating current, could be sent down the wire and the circuit completed through sea water. Oppositely phased current was led to two contacts through two silicon diodes. One contact (the speed contact) was fixed to the current meter; the other (the direction contact) was attached to a magnetic compass bar and could rotate freely.

The usual procedure in taking measurements aboard the *Horizon* was, when within good radar range (about 1000 m from the anchored buoy), to lower the Roberts meter to a depth of 300–330 m. The meter was allowed to come to equilibrium before taking any readings. More wire had to be let out if the instrument started to rise. Readings inadvertently made while the meter was rising or sinking rapidly were discarded.

Observations were made at depth intervals of 10 to 25 m. The velocity at each depth was recorded for a period of 20 to 150 sec. About every ten minutes the radar range and bearing from the anchored buoy was checked. The time for such a series of observations was between one and two hours. Often a free-floating surface drogue was released next to the anchored buoy before a Roberts meter lowering and picked up on its completion. In this way the surface current was measured.

Estimate of error in Roberts meter measurements

There are several sources of error in current measurements with this technique; some can be estimated with confidence, some cannot. It is believed that the errors are random, and that the various types of error are independent of one another. For those errors which can only be estimated, the values in the following discussion are not the outside limits of error but rather of the standard deviation.

Vol 6 1959

One type of error is introduced in assuming that the anchored buoy is a fixed point. The buoy may move, and this movement can be of two sorts; translational movement caused by dragging the anchor, and random movement because of the swinging of the buoy on its anchor. Movement due to anchor dragging was negligible. Except for the buoy at the equator which dragged five miles during the first 78 hours (less than 0·5 cm/sec), there was no movement of any of the anchored buoys large enough to be detected by celestial navigation, i.e., greater than a mile.

The length of wire between the underwater float and the surface float was in all cases 335 m. The depth of the underwater floats varied between 165 and 225 m. Assuming for a moment no horizontal displacement of the underwater float, the extreme diameter of the circle described by the buoy as it swings on its anchor would be the order of 600 m. Since the line between the surface float and the underwater float is curved rather than straight the circle would be less. On the other hand, any horizontal forces on the underwater float would increase the effective radius, perhaps by a factor of three.

Although no measure of this random buoy movement was taken on the *Dolphin* Expedition, measurements were made with a similar buoy in the Equatorial Countercurrent two months later. Bathymetric sounding lines were run with reference to the buoy. Then by matching different bathymetric surveys, the shift in the buoy position between surveys was determined. Nine such surveys were made around the buoy in a 17-day period, the first being run 36 hours after the buoy was laid. The anchor dragged for the first 150 hours at a speed of about 0·25 cm/sec. There was no net displacement for the remaining ten days. The seven positions of the buoy during this ten-day period fell within a circle of 700 m in diameter.

There were no rapid changes in current or wind observed during the current measurements. For an estimate of the buoy velocity, one might assume a movement of 700 m in six hours. The buoy velocity would be 3 cm/sec.

A second source of error is in the radar ranges. Under the ideal conditions which prevailed during most of the measurements, the precision of reading the Navy-type SU radar used was about $\pm$ 100 m and $\pm$ 1°. If one assumes that the ship drifted at a constant velocity during the 90 to 120 minutes it takes for a complete set of measurements, the error in measuring ship drift is about 2 cm/sec. Radar readings were taken every ten minutes, plotted and a smooth curve drawn through the points. Usually the ship drift appeared to be steady, but at times it was obvious that this varied. All estimates of ship velocity were made by fitting smooth curves through the plotted points. For the data taken aboard *Horizon*, it is believed that the ship drift relative to the buoy is known between $\pm$ 2 and $\pm$ 7 cm/sec.

A third source of error is in the current meters themselves. A speed calibration between 5 and 300 cm/sec was made for each instrument in the Convair tow tank, San Diego. Differences between runs with the same meter were as great as the differences between meters. Therefore, a single calibration curve was drawn for the three meters. The standard deviation was 1·5 cm/sec in the 5 to 50 cm/sec range based on 25 points and 7·5 cm/sec in the 50–300 cm/sec range based on 21 points. It is believed that this estimate of error is the largest to be expected. Difficulty was experienced in controlling the calibration at the high velocity end of the scale, and it is believed that at least some of the differences in the two standard deviations are a result of poor calibration control at higher velocities.

A direction calibration was made for each of the three meters, and again there was no significant difference between meters. The standard deviation from the true direction for the three meters was $\pm$ 8° based on 280 readings.

There are two other sources of error which are more difficult to estimate. One is ship yaw. As the ship is blown through the water with the wind on the beam, the ship's head swings and the ship begins to sail through the water with the wind abaft the beam. The ship continues to move with increased speed until the resistance of the water becomes too great; then the ship's head falls off, and the process is repeated. A ten-knot wind will cause a variation in *Horizon's* drift rate of 10–15 cm/sec. The period of this yaw is in the order of 20 minutes, which is too short to be detected by radar fixes but probably long enough for a full response of the current meter. It is perhaps possible to minimize this effect by throwing the rudder hard over, so that the ship will sail more easily and come back with the wind on its beam less readily; or steerage way and a constant heading can be maintained by steaming slowly through the water, as was done during most of the second half of the expedition.

The last source of error is the pumping action of the ship as it rides in the sea way causing the meter to ride up and down in the water and further causing it to swing somewhat from side to side. This motion causes the meter to record a higher apparent velocity than it should. It is difficult to guess how big a factor this is, but it must decrease as the amount of wire out increases.

For the second half of the expedition, and for some observations on the first half, the reference point was not a taut wire mooring but a free-floating parachute drogue with the parachute suspended at about 1000 m. Observations of such a drogue system placed near an anchored buoy indicated a westward movement of a few centimetres per second. Measurements made with respect to the free-floating parachute drogues necessarily have a somewhat greater uncertainty than those made with respect to the taut wire anchored buoy.

Ignoring the error caused by the pumping action of the ship and the error due to the drift of a parachute drogue, and summing up the rest of the errors, one can arrive at an estimate of the total error by taking the sum of the squares of the values listed previously. Thus, one arrives at a standard deviation between 11 and 18 cm/sec, depending upon the values used. It is believed that not many errors will exceed $\pm$ 30 cm/sec.

As an independent method of estimating the error all the Roberts meter observations from depths between 300 and 365 m were plotted from stations where anchored buoys or deep parachute drogues were used. As expected, the scatter of points for the parachute drogue stations is greater than for the anchored buoys. The average current from the anchored buoy stations is 7·5 cm/sec toward 036 °T, comparable with that determined by a Swallow current float, Table 2. The standard deviation of the vector distance from this mean value to the observed points was 14 cm/sec. For those stations made in reference to the deep parachute drogues the average current was 6·8 cm/sec toward 084 °T; the standard deviation from this average current was 24 cm/sec.

One is not justified in assuming that the current between 300 and 365 m was the same for all 57 stations. The scatter, therefore, reflects not only the errors in the system but changes in the current, and the standard deviation of the error in the Roberts meter must therefore be less than that indicated. On the other hand, the error due to pumping action must be small at this depth, while it may be significant in the top 10 to 20 m.

For the deeper part of the Cromwell Current between 300 and 1000 m, where velocities were too small to be measured with any precision by the modified Roberts meter, observations were made with Swallow current floats (SWALLOW, 1955).

THE CROMWELL CURRENT AS DEFINED BY DIRECT CURRRENT OBSERVATIONS

Representative Observations

The Cromwell Current or Pacific Equatorial Undercurrent is a thin, swift flow, centred on the equator and moving east beneath the west-flowing South Equatorial Current. Its core at 140 °W is at 100 m, where velocities as high as 125–150 cm/sec were recorded (51 cm/sec equals one knot). At 350 m the eastward flow was reduced to 10 cm/sec. At the equator the depth of the change from west to east flow is at about 20 m. The northern and southern limits of the current appear to be at about 2 °N and 2 °S respectively.

The north edge of the South Equatorial Current at the time of the five stations illustrated in Fig. 2 was south of the equator. Appreciable westward components were only found at the two stations at 1 °S and 2 °S. Occasionally a station in the Cromwell Current showed a marked north-south component; however, there appeared to be no consistent pattern to these transverse components.

The high velocity core (Fig. 2) is not very thick and the shear both above and below it is large. Changes of velocity of 150 cm/sec in 70 m or 2×10^{-2}/sec were common. The velocity gradient above the core is nearly always larger than that below it. The upper shear in the east-west component at stations taken along the equator was never less than $1{\cdot}5 \times 10^{-2}$/sec and never as high as 4×10^{-2}/sec.

Vol 6 1959

Cross Sections and Transport

Four cross sections of the current were made between 8–27 April along 140 °W. These are similar except in detail (Fig. 3). The Cromwell Current appears to be nearly symmetrical about the equator. Its jet-like appearance is misleading (Fig. 3)*; the current more closely resembles a thin ribbon. As defined by the 25 cm/sec contour,

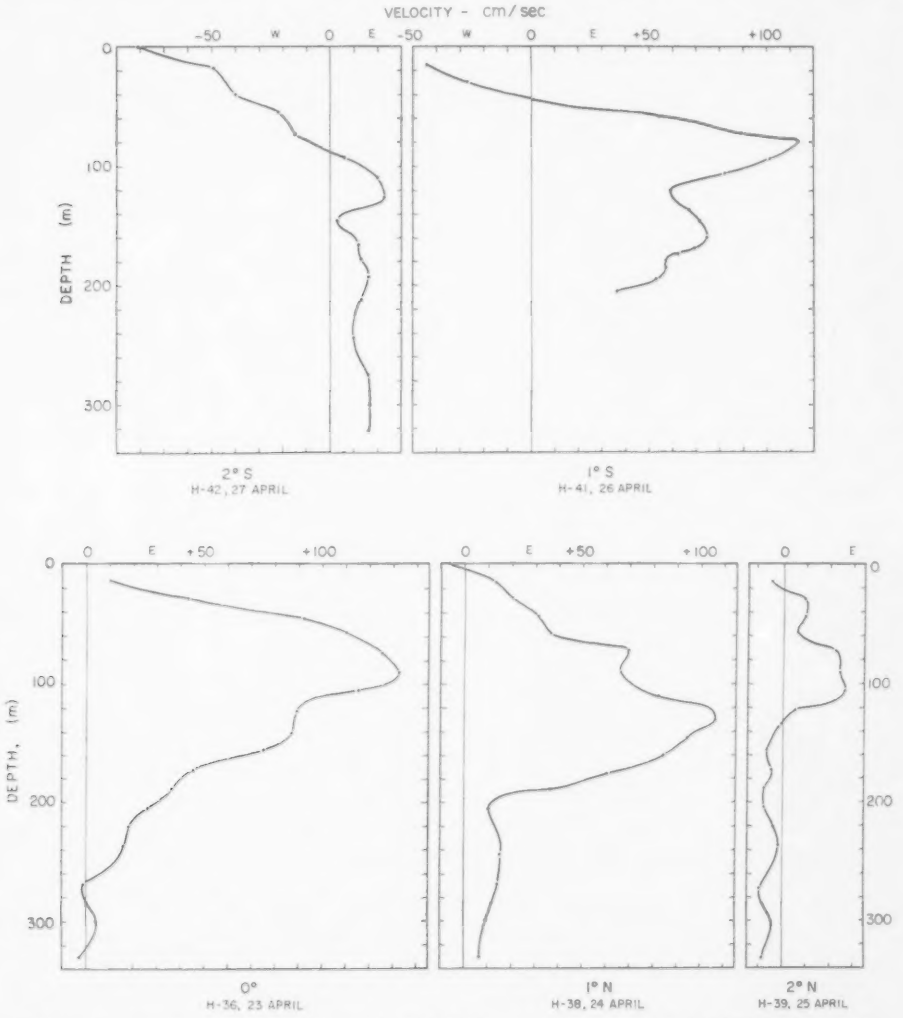

FIG. 2. East-west current profiles, made with Roberts current meter at 140 °W 25–27 April, Section IV.

it is about two-tenths of a kilometre thick and 300 km wide. The amount of water transported by the current is remarkably high, i.e. between 34 and 42 $\times$ 10^6 m^3/sec with a mean value of 39 $\times$ 10^6 m^3/sec (Table 1)†.

*The vertical exaggeration (Fig. 3) is 2 $\times$ 10^4.

†Comparable values for other confined, high-velocity currents are the Gulf Stream 74 $\times$ 10^6 m^3/sec; Kuroshio, 65 $\times$ 10^6 m^3/sec; and the Florida Current 26 $\times$ 10^6 m^3/sec (SVERDRUP *et. al.*, 1942).

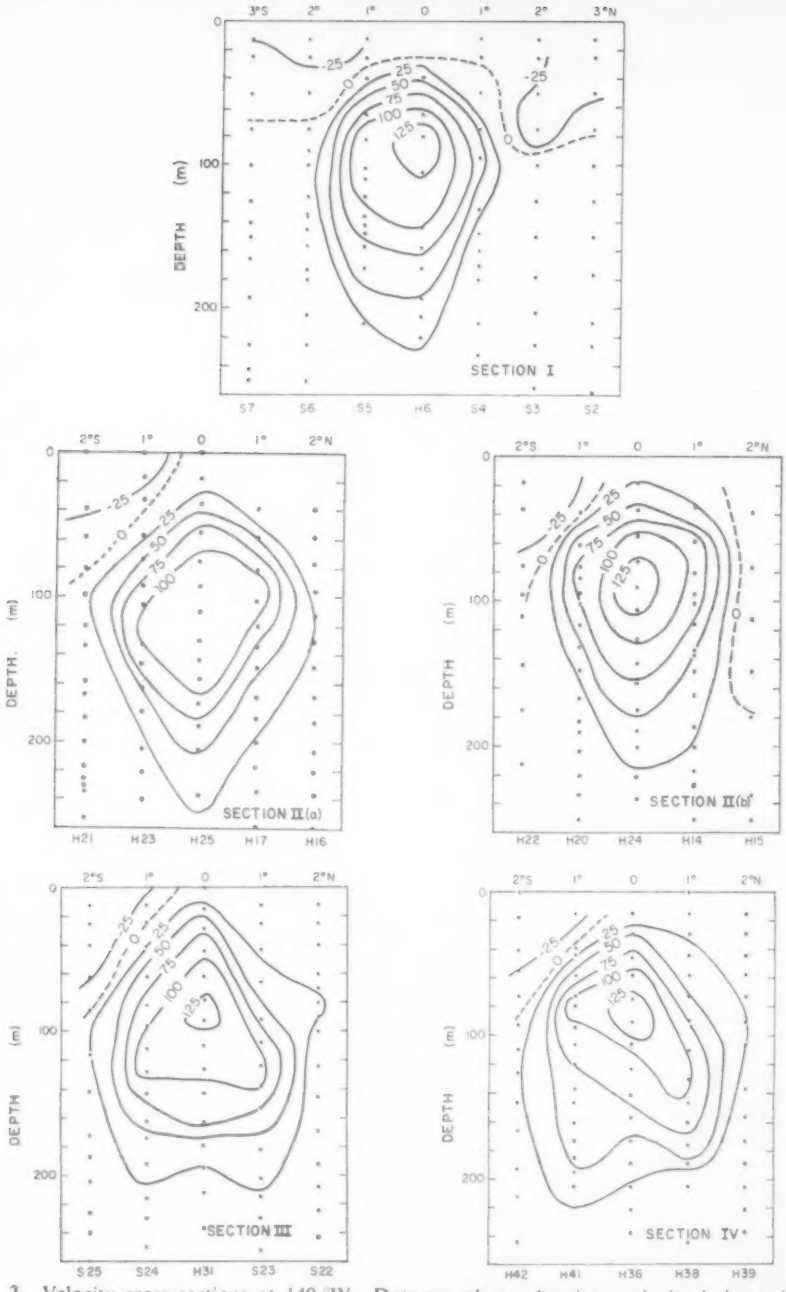

FIG. 3. Velocity cross sections at 140 °W. Dots are observed points, velocity is in cm/sec; plus is eastward and minus is westward flowing current; station numbers are along the bottom. Section I (6–9 April) : all but the equator station were made by *Hugh M. Smith* with reference to deep parachute drogues. Section II (12–18 April) : at least two sets of measurements were made by *Horizon* at each anchored buoy. Section II (a) is drawn from data which give the largest transport; Section II (b) is drawn from data which give the smallest transport. Section III (20–22 April) is drawn from data obtained by *Hugh M. Smith*. Section IV (23–27 April) was made by *Horizon*. The stations are the same as those shown in Fig. 2.

Vol. 6 1959

Swallow current floats were used to measure the low velocities at depths greater than 250 m. Four successful measurements were made (Table 2). At 300 m the eastward flow was only 11 cm/sec; at 500 m the direction was reversed; at 700 m it was moving westward at 13 cm/sec, but was again weak at 1000 m. Thus, the Cromwell Current is embedded between two westward-flowing currents (Fig. 4).

Table 1. *Cromwell Current mass transport at* 140 °*W*

Date	*Ship*	*Section*	*Transport* (m^3/sec)
April 6– 9	*Hugh M. Smith*	I	34×10^6
April 12–18	*Horizon*	IIa	42×10^6
April 12–18	*Horizon*	IIb	34×10^6
April 20–22	*Hugh M. Smith*	III	42×10^6
April 23–27	*Horizon*	IV	42×10^6
			39×10^6 Average transport

Table 2. *Current velocities from Swallow floats*

Date	*Loaded Depth* (m)	*Calculated Depth* (m)	*Hours Tracked*	*Speed* (cm/sec)	*Direction* (*toward*)
April 21	700	670 ± 210	10	14 ± 1	285°
April 22	500	440 ± 125	26	7 ± 0·5	290°
April 23	300	420 ± 135	10	12 ± 1	075°
April 26	1000	1020 ± 365	18	2 ± 0·7	205°

Steadiness of the Cromwell Current

From 11–24 April, 32 usable Roberts meter stations were made on the equator at 140 °W. There was no significant change in the velocity structure during this period (Fig. 5). The peak velocity varied between 120 and 150 cm/sec. Its depth (dotted line, Fig. 5) remained rather constant at 100 ± 12 m.

These results are in marked contrast to the Equatorial Countercurrent which flows eastward along the surface 500 miles to the north of the Cromwell Current. Recent observations there suggest that the Countercurrent can change direction ninety degrees, stop or double its speed all within a three-day period (KNAUSS, unpublished results).

The Eastward Extent of the Cromwell Current

The Cromwell Current was found at all stations from 140 °W to 92 °W inclusive. Sixteen miles east of the 92 °W station and straddling the equator is Isabella Island, the largest island in the Galapagos. The Cromwell Current was weak, but present, sixteen miles west of this island, but it was not found at the single station east at 89 °W. As pointed out in the introduction, Cromwell did not find the current at 84 °W east of the Galapagos.

Velocities in excess of 100 cm/sec are found at all stations through 95 °W. Between 140 °W and 95 °W the core (dotted line, Fig. 6) becomes shallower towards the east. The mean slope is $1{\cdot}3 \times 10^{-5}$; however, most of the slope occurs between 140 °W and 116 °W. Between 95 °W and 92 °W the current definitely weakens, and the core slopes downward. What current there is at the station sixteen miles to the west of

Isabella Island is between 80 and 100 m. Except at the eastern terminus of the current, the variation in the velocity profile with longitude is no greater than that observed during the 18 days at the equator, 140 °W. The South Equatorial Current occasionally reached as far north as the equator (131 °W and 122 °W), but its presence appeared to have little effect on the Cromwell Current.

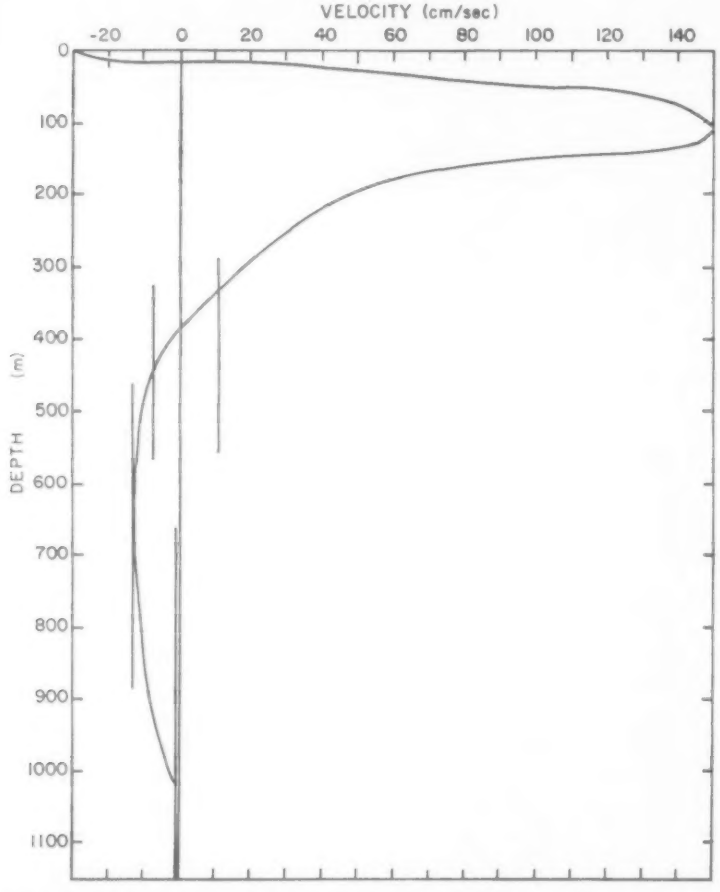

FIG. 4. A composite east-west velocity current profile, 0–100 m, at the equator at 140 °W, based on Roberts current meter and Swallow current floats. Uncertainty in depth of Swallow floats is indicated by vertical lines.

COMPARISON OF THE CROMWELL CURRENT WITH THE HYDROGRAPHIC OBSERVATIONS

One of the objectives of *Dolphin* was to obtain hydrographic measurements to correlate with the Cromwell Current, with the hope that hydrographic observations would provide some insight into the nature and cause of the current. Above the strong thermocline of the eastern tropical Pacific is a layer of warm, well-mixed, almost isothermal water, 10 to 100 m thick with typical temperature values of

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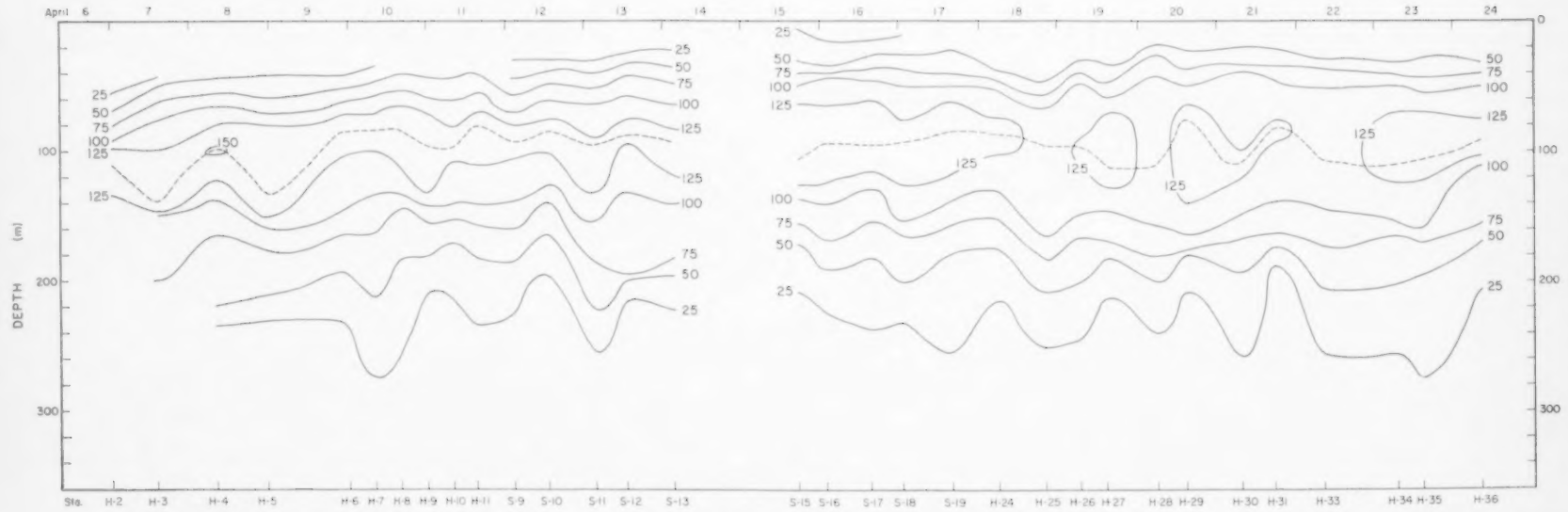

FIG. 5. Variation in east-west velocity component at the equator. Velocity is in cm/sec. Stations are listed along the bottom. The dashed line represents the depth of maximum velocity.

Vol
6
1959

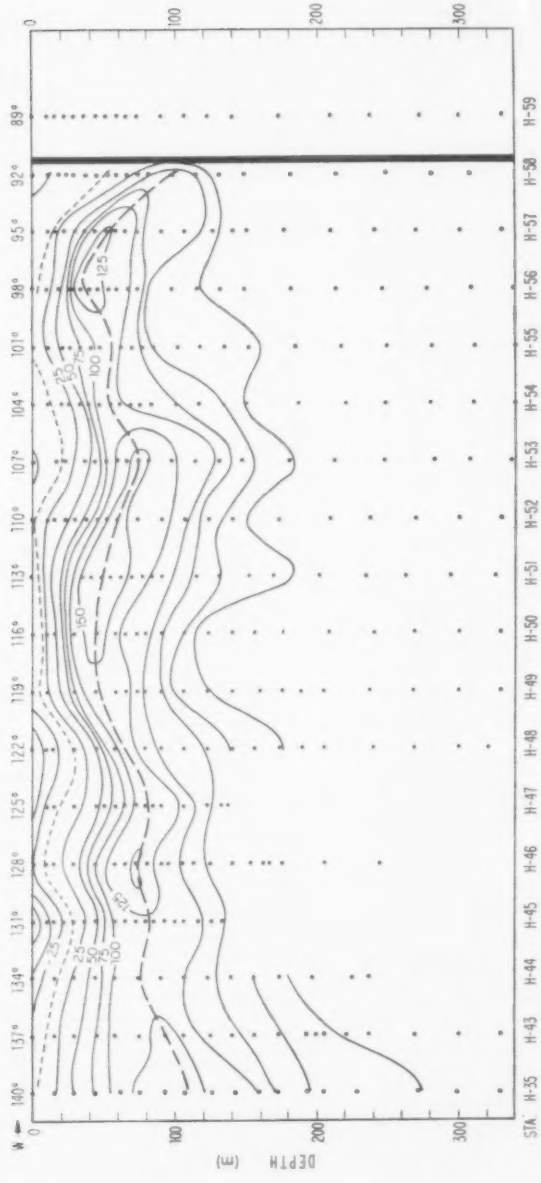

FIG. 6. East-west velocity components along the equator. Stations were made once a day. Dots are observed points. The dashed line is the depth of maximum velocity.

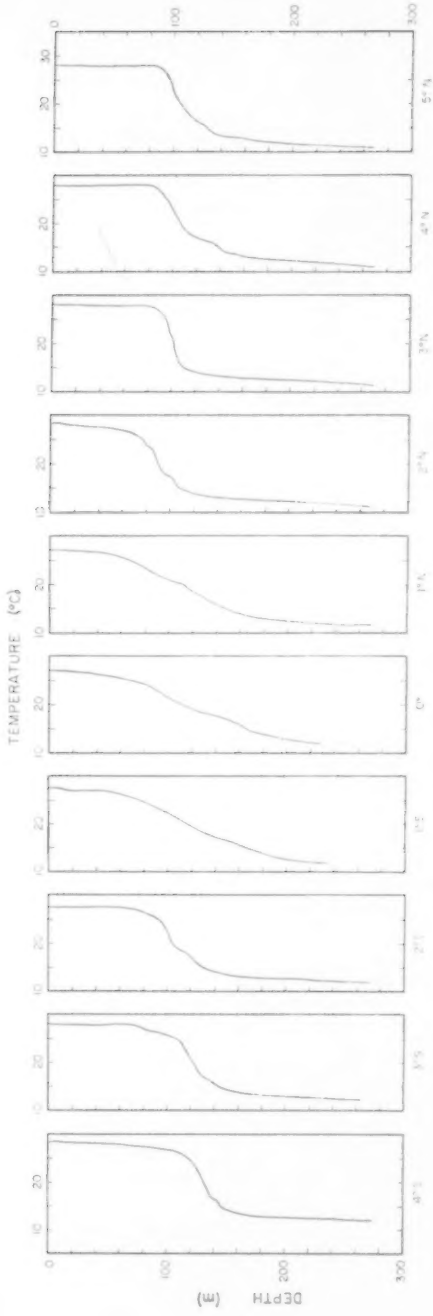

FIG. 7. Bathythermograms, 5 °N–4 °S along 140 °W.

Vol
6
1959

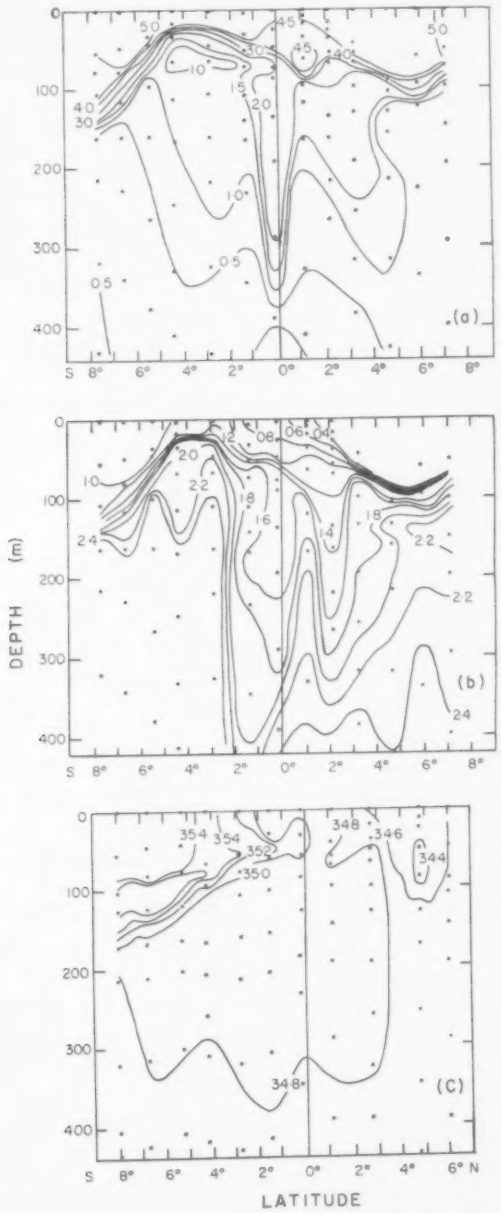

FIG. 8. Cross sections of (a) oxygen, 110 °W; (b) phosphate, 110 °W; (c) salinity, 120 °W (oxygen is in millilitres/litre and phosphate is in microgram atoms/litre). Redrawn from data in KING *et al.*, (1957).

25°–28 °C. Just below the mixed layer the temperature can decrease 10 °C within 25 m in some regions. Below the thermocline at 300 m the temperature is 10°–12 °C.

The warm surface layer over a strong thermocline occurred on both sides of the equator on *Dolphin*, but not at the equator (Fig. 7). It would appear that strong vertical mixing takes place in the top few hundred metres. The breakdown of the sharp temperature discontinuity can be seen in the spreading of the isotherms near the equator. Similar changes in distribution can also be seen in the distribution of salinity, oxygen, phosphate and silicate (CROMWELL, 1953; WOOSTER and CROMWELL, 1958; KING *et. al.*, 1957). The eastern tropical Pacific is characterized by a marked oxygen minimum just below the thermocline. At the equator, however, water of high oxygen content is mixed down from the surface to depths of at least 300 m; near-surface values there are lower than on either side. Similarly, water low in phosphate appears to be mixed downward through the thermocline. The tongue of water with a high salinity which moves up from the south along the thermocline does not cross the equator. Presumably this maximum is dissipated by mixing (Fig. 8).

The colder surface water and the termination of the saline tongue at the equator are suggestive of upwelling (CROMWELL, 1953). Presumably the south-east trades would cause a divergence at the equator. From the Ekman wind drift theory, winds with an easterly component would create a northward drift, north of the equator and a southward drift, south of the equator (CROMWELL, 1953; NEUMANN, 1947). The anomalous near-surface distributions (low temperature and oxygen, high phosphate and silicate, and lack of salinity maximum) have been used as criteria for upwelling (WOOSTER and CROMWELL, 1958; AUSTIN and RINKEL, 1957), but do not explain the analogous situation below the thermocline where high temperature and oxygen and low phosphate and silicate values are found. It had been suggested that this 'trough' in the isopleths might be related to the Cromwell Current (WOOSTER and JENNINGS, 1955).

No one seems to have suggested previously that the anomalous distribution of properties in the region of the equator might all be related to a single process—strong vertical mixing at the equator. Similar features can be seen on nearly every meridional section between 90 °W and 180 °. Based on the *Dolphin* observations, the north-south boundaries of the Cromwell Current appear to agree approximately with the limits of mixing indicated by temperature and oxygen data (Fig. 9).

THE CROMWELL CURRENT AS A GEOSTROPHIC CURRENT

The hydrographic data suggest that the water in the upper few hundred metres near the equator is rather well mixed. The sharp vertical density gradient over most of the eastern tropical Pacific that separates high temperature surface water from low temperature water (and acts as a boundary between the surface water which has high oxygen and low phosphate content and the deeper water which is deficient in oxygen but rich in phosphate) is reduced in magnitude at the equator because the gradient is extended over a greater range of depth. Mixing carries water characterized by high temperature, high oxygen and low phosphate to greater depths at the same time that water of lower temperature, lower oxygen and higher phosphate is brought into the surface layer. However, the surface acts as a heat and oxygen source and a phosphate sink (the latter because of increased photosynthesis). For this reason the

Vol 6 1959

horizontal temperature, oxygen and phosphate gradient across the equator is less marked at 40 m than at 200 m (Fig. 8–9).

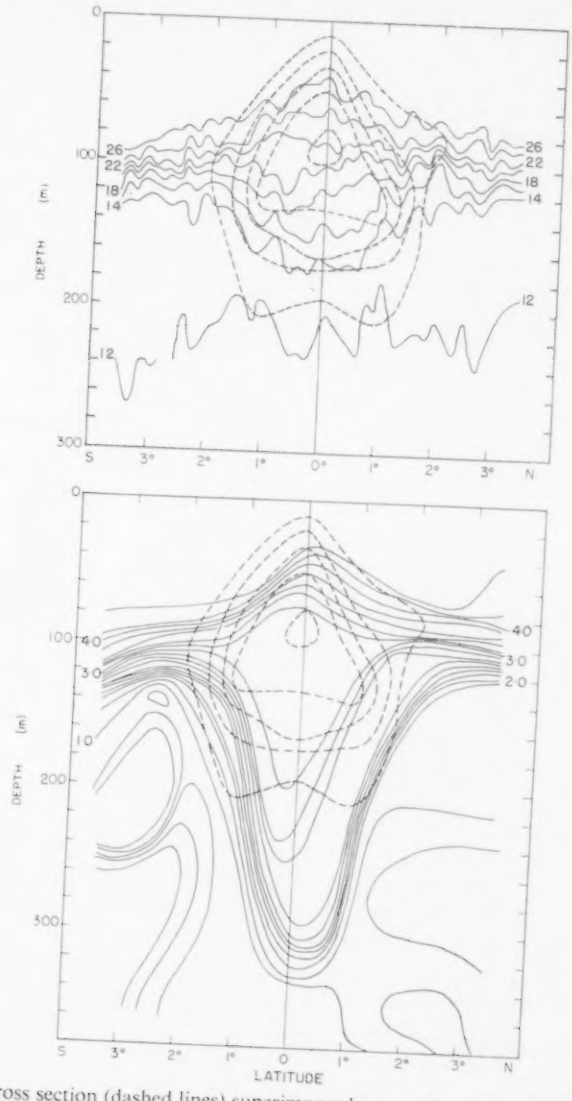

FIG. 9. Velocity cross section (dashed lines) superimposed on temperature and oxygen sections. Velocity contours 25–125 cm/sec identical with Fig. 3, Section III. Temperature is in degrees centigrade, oxygen is in ml/litre. All data taken by *Hugh M. Smith* with Roberts meter, bathythermograph and hydrographic stations on Section III, 140 °W, 20–22 April.

The Cromwell Current is in approximate geostrophic balance (that is, the balance of forces in the north-south direction, except very near the equator is represented by the Coriolis force and the pressure gradient, $fu = \alpha\, \partial p/\partial y$ where u is the eastward velocity, α is the specific volume of sea water, $\partial p/\partial y$ is the north-south pressure gradient, and $f = 2\,\omega \sin \phi$ where ω is the angular rotation of the earth and ϕ is the latitude).

It is suggested that the meridional component of pressure gradient necessary for a geostrophic Cromwell Current is the result of vertical mixing at the equator (see discussion below). To understand how this mixing could result in an eastward subsurface flow, let the density structure of the ocean be represented by a simple two-layer system of constant thickness and imagine this system to extend across the equator. There would be no north-south horizontal pressure gradient under these circumstances and therefore, no geostrophic current. Now, mix the water at the equator. The resulting density structure at the equator and the geostrophic current at 1° are shown in Fig. 10; the unmixed water is at 2°, and it is assumed that the

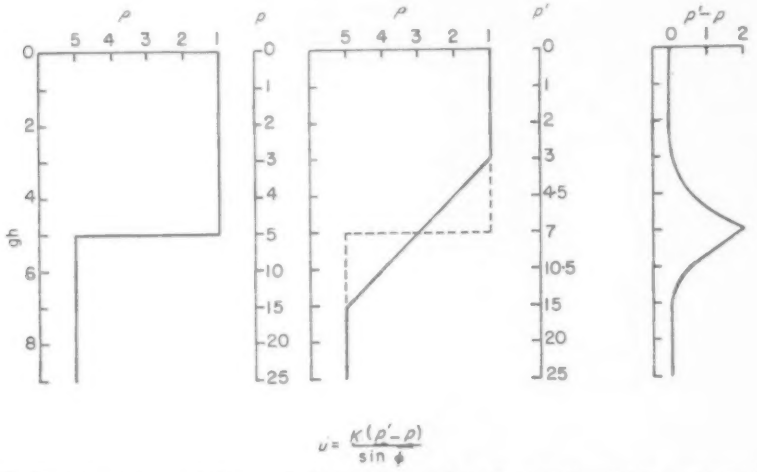

FIG. 10. Two-layer system before and after mixing, and resulting horizontal pressure gradients. $p = \int \rho g \, \partial h$. ρ and gh given in arbitrary units.

pressure gradient at any given depth is constant between the equator and 2°. The pressure gradient changes sign at the equator, as does the sine of the latitude, so that there is no change in the direction of the current. The current is symmetrical about the equator.

The resulting geostrophic current at 1° can be calculated from a more realistic model of the density structure before and after mixing, again assuming that there is no mixing at 2° and a constant gradient between 2° and the equator (Fig. 11). This density model is in excellent agreement with the observed density (Fig. 12). The geostrophic current is of the correct order of magnitude; the core is at approximately the correct depth, and the shape of the current is similar to that observed (Fig. 12). This model is consistent with the rise in the core of the Cromwell Current to the east. If the Cromwell Current is in geostrophic balance, and if the pressure gradient is the result of mixing across the thermocline, then one would expect the core to become shallower to the east in agreement with the shoaling of the depth of the thermocline to the east (AUSTIN, 1958).

The Coriolis parameter approaches zero near the equator, and the question arises as to whether or not one can expect the other terms in the equation of motion to

Vol 6 1959

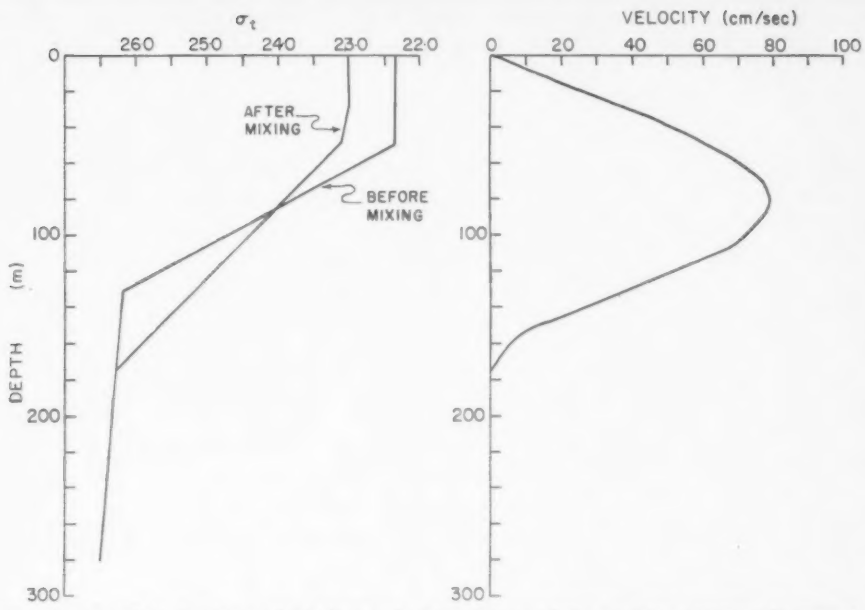

FIG. 11. Realistic density model before and after mixing and the resulting geostrophic current. It is assumed the unmixed water is at 2°, and the mixed water at the equator, and the pressure gradient constant between 0° and 2°. The geostrophic current has been calculated for 1°.

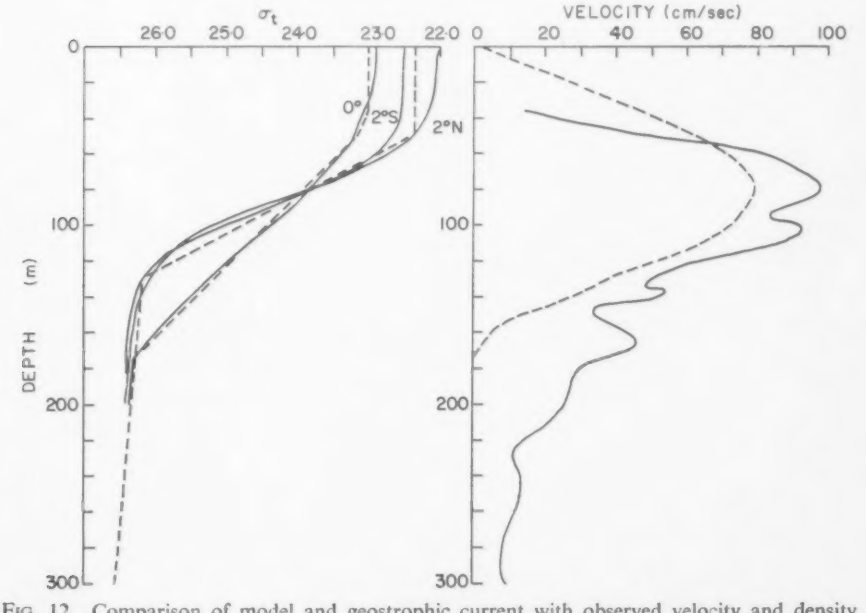

FIG. 12. Comparison of model and geostrophic current with observed velocity and density distribution. The dashed lines are from Fig. 11. The solid lines of σ_t are drawn from hydrographic data along 140° W at the latitudes indicated. The velocity profile is from a Roberts meter station at 1° N.

be small enough to allow the pressure gradient to remain approximately in balance with the Coriolis term. In the meridional plane the equation is:

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} = -\alpha\frac{\partial p}{\partial y} - uf + \alpha A_z\frac{\partial^2 v}{\partial z^2} + \alpha A_h\left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2}\right)$$

where A_z and A_h are vertical and lateral eddy coefficients respectively. The values of uf for $0{\cdot}5^\circ \leqslant \phi \leqslant 2^\circ$ (Table 3) are for a value of u comparable to those observed at 140 °W at a depth of 100 m.

Since little really is known about the nature of the meridional circulation it is not possible to calculate terms other than the pressure gradient with any degree of certainty. However, TSUCHIYA (1955) has shown that if one uses representative values of $v = 5 - 10$ cm/ sec, $A_z = 10^2$ gm/cm/sec and $A_h = 10^7$ gm/cm/sec, the values of the other terms are all less than 10^{-5} cm/sec^2. This is an order of magnitude less than the Coriolis term corresponding to the observed velocity (Table 3).

For other evidence that geostrophic balance might be expected when the value of the Coriolis term is the order of 10^{-4} cm/sec^2, it can be pointed out that comparisons in the Equatorial Countercurrent at 5–8 °N of GEK surface current measurements with those calculated from the geostrophic equation are in good order first agreement. The magnitude of the terms is $5-8 \times 10^{-4}$ cm/sec^2, only two to four times larger than those given in Table 3.

Table 3. *Values of Coriolis terms near equator using representative velocities measured at* 100 *m. The pressure gradient in the last column is that needed for geostrophic balance.*

ϕ	sin ϕ	u (cm/sec)	uf (cm/sec^2)	$\frac{\Delta h}{\Delta y}\left(\frac{\text{dynamic cm}}{\text{degree of latitude}}\right)$
0·5	0·009	125	$1{\cdot}6 \times 10^{-4}$	1·8
1·0	0·017	85	$2{\cdot}0 \times 10^{-4}$	2·2
1·5	0·026	50	$1{\cdot}9 \times 10^{-4}$	2·1
2·0	0·035	25	$1{\cdot}3 \times 10^{-4}$	1·5

It is difficult to make a detailed comparison of the geostrophic current and the Cromwell Current using the *Dolphin* data. Differences of one to two dynamic centimetres per degree of latitude are all that are required (Table 3). However, differences of two to four dynamic centimetres were observed on the equator station over a six-day period. This kind of variation in the observed pressure tends to obscure any meridional gradient observed on stations taken on different days. However, the tendency for a pressure gradient to be symmetrical about the equator can be seen in Fig. 13. The necessary pressure gradient of two dynamic centimetres per degree of latitude is approximately realized. A similar picture of the pressure gradient can be seen on the plot of dynamic height at 100 m relative to 1000 m, but the scatter of values is larger. The argument for using the 400 m surface as the 'level of no motion' is the current reversal at about 400 m (Fig. 4).

It should be noted that the geostrophic balance applies only to the meridional forces; the nature of the balance of the zonal forces is yet to be determined. It should also be noted that the models in Fig. 10 and 11 can be adjusted to include a westward

flowing South Equatorial Current at the surface. One need only begin with a mixed layer which is shallower at the equator than at 2°.

From the foregoing, the Cromwell Current is in approximate geostrophic balance. The horizontal pressure gradients appear to result from vertical mixing of the water at the equator, and a breakdown of the sharp tropical thermocline. The mixing creates a meridional pressure gradient symmetrical about the equator; hence, the Cromwell Current is a subsurface eastward flow which balances this gradient. All of the observational data support this idea and the magnitude of the different force terms suggests that a balance between the pressure gradient and the Coriolis term is to be expected even to within half a degree of the equator.

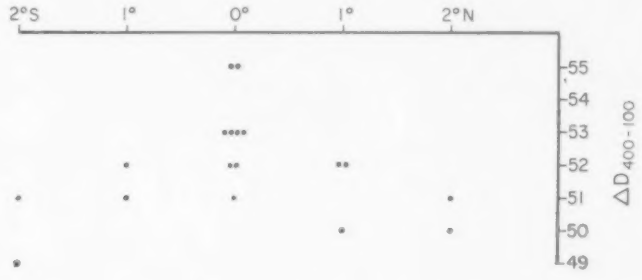

FIG. 13. Distribution of values of dynamic height ($\Delta D_{400-100}$) at 100 m relative to 400 m along 140 °W at all *Horizon* hydrographic stations. Uncertainty of pressure gradients of 1 dynamic centimetre/degree latitude at 1° is equivalent to an uncertainty of 40 cm/sec in the resulting geostrophic current.

The fact that the Cromwell Current is a geostrophic current, does not, of course, explain its existence, nor is this explained by the meridional pressure gradient which appears to result from strong vertical mixing at the equator; because it can be argued that strong turbulence is associated with the high velocities of the current. Thus the Cromwell Current could be the cause of mixing, rather than the result. Hence, one must find a mechanism other than the current itself that will create strong mixing at the equator to depths of 300 m, but not on either side of it.

Little is known or understood about upwelling and vertical mixing processes in the ocean, but it is generally believed that upwelling is a comparatively shallow process and does not extend as deep as 300 m. However, the Cromwell Current might be produced if upwelling created by the wind-induced divergence at the equator could cause vertical mixing to depths of 300 m.

DISCUSSION AND CONCLUSIONS

(1) The observations of the *Dolphin* Expedition have shown that the eastward flow beneath the surface at the equator (CROMWELL *et. al.*, 1954) is a current of major proportions. These data indicate that the Cromwell Current is at least 3500 miles long with a transport at 140 °W of 35–40 $\times$ 10^6 m^3/sec. This transport is greater than hitherto estimated for any current in the equatorial Pacific and, with the exception of the Kuroshio, for any current in the Pacific (SVERDRUP *et. al.*, 1942).

The existence of a new current with a transport of this magnitude, however, suggests that transport balance in the equatorial Pacific should be re-examined. Further evidence that previous transport estimates are incorrect was found four months later on *Doldrums* (the third IGY Expedition made by the Scripps Institution of Oceanography). Current measurements with the modified Roberts meter and Swallow floats at 7° 52′ N, 107° 30′ W suggested that the transport of the Equatorial Countercurrent is at least twice as great as previously estimated (KNAUSS and PEPIN, 1959). The Cromwell Current and the increase flow in the Countercurrent more than triple the total eastward transport in the equatorial Pacific as given by SVERDRUP *et. al.* (1942).

(2) The length of the Cromwell Current can be inferred by examining hydrographic data.* All available hydrographic sections east of 170 °W show evidence of the current, but none of those examined from the west of 160 °E suggest its presence; some sections between 170 °W and 160 °E are indicative of it. From somewhat similar criteria it appears that ' . . . . . upwelling is usually of a comparatively minor nature or may be absent west of 180° ' (AUSTIN and RINKEL, 1957).

However, recent Japanese current measurements in January–February, 1958, suggest the Cromwell Current at 150 °E (YOSIDA *et. al.*, 1959). Their temperature sections also show the characteristic breakdown of the thermocline at the equator.

If the Cromwell Current is a response to the pressure field produced by the wind-induced divergence at the equator, then the easterly component of the south-east trades is a measure of the driving force. The current should be weak or missing east of 180°, since the streamline charts of the mean surface winds for January and July (MINTZ and DEAN, 1952) show no easterly component to the tradewinds along the equator west of 180° during January and only a weak component west of 170 °E during July. The question of the nature and continuity of the Cromwell Current in the western Pacific must be investigated.

(3) Perhaps the biggest problem posed by the data is where does the water go. One possibility is that it is lost to the sides as the current moves downstream. The fact that the core velocity at the equator showed little change between 140 °W and 95 °W does not necessarily imply that the transport is the same throughout this distance. A decrease in the current strength east of 140 °W is suggested by the Mintz-Dean streamline charts of surface winds, as there is a decrease in the easterly component of trades along the equator east of 130 °W. At 90 °W there is no easterly component at all in summer and only a very weak one in winter.

If, on the other hand, the transport is still high at 95 °W, then the current must be deflected either north or south of the equator by Isabella Island. Since water of the Cromwell Current is characterized by high oxygen values, the oxygen content is an excellent tracer. The oxygen distributions on *Shellback* Sections X and XII (July and August, 1952) suggest that the current may be found on both sides of the Galapagos at 91 °W and at the equator at 85 °W (WOOSTER and CROMWELL, 1958). If the *Shellback* oxygen data mean that the current is present east of the Galapagos

*One might hope to find evidence of a similar flow in the Atlantic by examining the hydrographic data; however, there are only six meridional sections across the equator in the Atlantic. *Meteor* Profiles IX at 30 °W, Xa at 10 °W and Xb at 0° show evidence of vertical mixing in the oxygen section (WATTENBERG, 1939). *Atlantis* sections at 30 °W, 34 °W and 38 °W (*Bulletin Hydrographique*, 1952, Copenhagen) do not show any clear evidence for vertical mixing at the equator, but it appears possible that such evidence could be obscured because the number of observations in the top 300 m are fewer than on the *Meteor* Expedition.

in 1952, then either it is not present at all times, or else it is so weak that it was not detected by the parachute drogues used by Cromwell at 84 °W in November, 1955, or by the Roberts meter on *Dolphin* at 89 °W in June, 1958). Perhaps water is entrained as it moves through the islands; hence, there may be only a slow eastward drift east of the Galapagos. This problem cannot be solved until a series of sections east of 140° W become available to determine the transport of the Cromwell Current at successive points along the stream.

(4) It seems likely that the Cromwell Current surfaces at times. Perhaps the most likely place for this is near the Galapagos, where Isabella Island may provide an effective block to further passage of the water and make the current structure unstable. Evidence for this can be found on the Deutschen Seewarte surface current chart which shows the current at the equator 300–600 miles east of the Galapagos to be either east or west in March (SCHOTT, 1943). This region is the only place on the chart where such a variation in the current direction is indicated.

(5) Upwelling at the equator must be re-examined in the light of the Cromwell Current. It may be that upwelling along the equator will be a necessary part of any explanation of the Current. However, if the current can be explained without resorting to upwelling, then one must question whether there is indeed any net upward flow of water at the equator. The evidence for it has been based primarily on the low surface temperatures there. However, as already pointed out, all the hydrographic data suggest that there is strong vertical mixing along the equator. Furthermore, upwelling *per se* cannot account for the high temperatures, high oxygen and low phosphate values beneath the expected depth of the thermocline. The colder surface temperature can be explained by vertical mixing (Fig. 11), and it seems likely that such mixing could be explained by the high velocities and high velocity shears recorded at the equator.

Acknowledgements—Without the assistance of the Pacific Oceanic Fisheries Investigation, the work reported here would be less complete. JOSEPH KING and MURICE RINKEL spent long hard hours aboard *Hugh M. Smith* working with unfamiliar equipment. RINKEL and THOMAS AUSTIN gave freely of their time in helping me analyze their results. The key instrument of this expedition was the modified Roberts current meter. It was developed by JAMES SNODGRASS and was used successfully throughout the cruise.

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Vol
6
1959

A review of the calculation of ocean currents at the equator

ROBERT S. ARTHUR

(*Received* 18 *August* 1959)

Abstract—Wind stress and horizontal pressure gradient at the equator have been used in the calculation of velocity. Various formulae are considered from the standpoint of consistency with an undercurrent. A simple relation between the east-west pressure gradient and the curvature of the velocity profile yields an undercurrent, but application is uncertain because of neglect of accelerations. The most promising formula appears to be that derived from the geostrophic approximation for zonal motion. Both friction and acceleration appear to be important in the dynamics of the Equatorial Undercurrent.

SEVERAL formulae for the calculation of ocean currents at the equator have appeared during the last decade. The purpose of this review is to examine such a calculation in respect to observations of the Equatorial Undercurrent, a narrow, eastward current prevalent in the lower part of the surface layer and in the thermocline layer along the equator in the Pacific. Among the important features observed along the equator are the following (CROMWELL *et al.*, 1954 ; KNAUSS and KING, 1958 ; KNAUSS, 1960) :

(a) surface flow may be directed westward at speeds of 25-75 cm sec^{-1}

(b) current reverses at a depth of from 20 to 40 m

(c) eastward undercurrent extends to a depth of 400 m with a transport of as much as 30×10^6 m^3 sec^{-1}

(d) core of maximum eastward velocity (50–150 cm sec^{-1}) rises from a depth of 100 m at 140 °W to 40 m at 98 °W, then dips down

(e) undercurrent appears to be symmetrical about the equator and becomes much thinner and weaker at 2 °N and 2 °S.

The various calculations of currents at the equator have been made by starting with the dynamic equations in some approximate form (WEENINK and GROEN, 1952; JERLOV, 1953 ; YOSHIDA, *et al.*, 1953 ; WYRTKI, 1956 ; HIDAKA and NAGATA, 1958). In most cases, a surface boundary condition (wind stress) has been introduced, and a bottom boundary condition (zero stress at some depth) has been assumed. The density distribution and resulting pressure distribution have been assumed to be known through observations of salinity and temperature at depth. This last assumption distinguishes the calculation methods from the theories of wind-driven circulation which include the equatorial region. In the theories (e.g., SVERDRUP, 1947 and REID, 1948a) the pressure distribution in integrated form is one of the results from the solution of the equations and is not assumed to be known.

The present paper directs attention primarily toward the calculation methods, but some comments on the theories are included. Except for the results of HIDAKA and NAGATA (1958), little or no indication of the Undercurrent has come from

previous applications of calculation formulae, and no clues to the existence of this current come from the theories. The situation is understandable in the case of calculations in which only a surface current is sought and in the theories which are involved with mass transport only. However, it seems appropriate to examine the bases of previous results to see whether an undercurrent is consistent with the equations and assumptions used.

As a basis for a review of the different methods, a set of dynamical equations for steady-state conditions is introduced.

$$u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z}=fv-2\Omega w\cos\phi-\frac{1}{\rho}\frac{\partial p}{\partial x}+\frac{1}{\rho}\frac{\partial \tau_x}{\partial z}+\frac{A_H}{\rho}\left(\frac{\partial^2 u}{\partial x^2}+\frac{\partial^2 u}{\partial y^2}\right). \quad (1)$$

$$u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+w\frac{\partial v}{\partial z}=-fu-\frac{1}{\rho}\frac{\partial p}{\partial y}+\frac{1}{\rho}\frac{\partial \tau_y}{\partial z}+\frac{A_H}{\rho}\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}\right). \quad (2)$$

$$u\frac{\partial w}{\partial x}+v\frac{\partial w}{\partial y}+w\frac{\partial w}{\partial z}=2\Omega u\cos\phi-g-\frac{1}{\rho}\frac{\partial p}{\partial z}, \quad (3)$$

$$\frac{\partial}{\partial x}(\rho u)+\frac{\partial}{\partial y}(\rho v)+\frac{\partial}{\partial z}(\rho w)=0. \quad (4)$$

The rectangular coordinate system has its origin at the sea surface, with x directed eastward, y northward, z upward, and u, v, w represent the corresponding velocity components. The Coriolis parameter is $f=2\Omega\sin\phi$, where Ω is the angular speed of the earth's rotation and ϕ is latitude. The pressure is p, the density ρ, the acceleration of gravity g, and the x- and y-components of shearing stresses across horizontal planes are τ_x and τ_y, respectively. Equations (1) and (2) represent the horizontal component equations of relative motion with the coefficient of lateral eddy viscosity, A_H, taken as uniform. The pressure will be assumed to be *known* from observations of density through the usual hydrostatic form of the approximation (3) of the vertical component equation and assumption of a reference surface. Assumption of a steady density field gives the equation of continuity in form (4).

A detailed analysis of velocity calculation at the equator is made from these equations in simplified form. The terms to be retained in the equations are those which have been suggested to be the important ones in central oceanic areas, i.e., beyond coastal boundary zones (MUNK, 1950 ; STOMMEL, 1956). The object of the analysis is to test the selection of terms against the new piece of observational evidence – the Undercurrent. An inquiry is pertinent in view of the fact that acceleration terms have been neglected at the equator in the calculation methods and theories, although FOFONOFF and MONTGOMERY (1955) have found acceleration to be important near the equator in their model of the Undercurrent.

It will be assumed that the stress and velocity at the surface are known. The applicability of the approximations will be examined, and comparison will be made with other calculations of currents at the equator.

ANALYSIS

STOMMEL (1956 ; 1957) proposes that, in general, the friction layer is limited to the proximity of the sea surface. Below this surface layer, the horizontal currents are approximately geostrophic, but there is vertical velocity which permits steady-state

Vol 6 1959

conditions for ρ. If acceleration, lateral friction, and the term $2\Omega w \cos\phi$ are neglected in (1)–(4), the horizontal convergence (or divergence) may be obtained without any lateral (coastal) boundary condition, and STOMMEL (1956) used such results to calculate v and w below the friction layer.

At the equator, the layer in which the friction or acceleration terms are important must be as deep as the layer in which the pressure terms are significant, according to (1) and (2). Since an observed distribution of ρ will be introduced, thermohaline forces as well as wind stress may be influencing the pressure gradient calculated from the hydrostatic form of (3). Relatively strong horizontal currents with important vertical shear may occur well below the depth to which the frictional layer is usually limited at higher latitudes. The possibilities for an undercurrent under the assumptions made by STOMMEL and others (e.g., SVERDRUP, 1947) may be examined with only very simple mathematics because at the equator ($\phi = 0°$ and $f = 0$) equations (1) and (2) may be written in the form

$$0 = -\frac{\partial p}{\partial x} + \frac{\partial \tau_x}{\partial z}, \tag{5}$$

$$0 = -\frac{\partial p}{\partial y} + \frac{\partial \tau_y}{\partial z}. \tag{6}$$

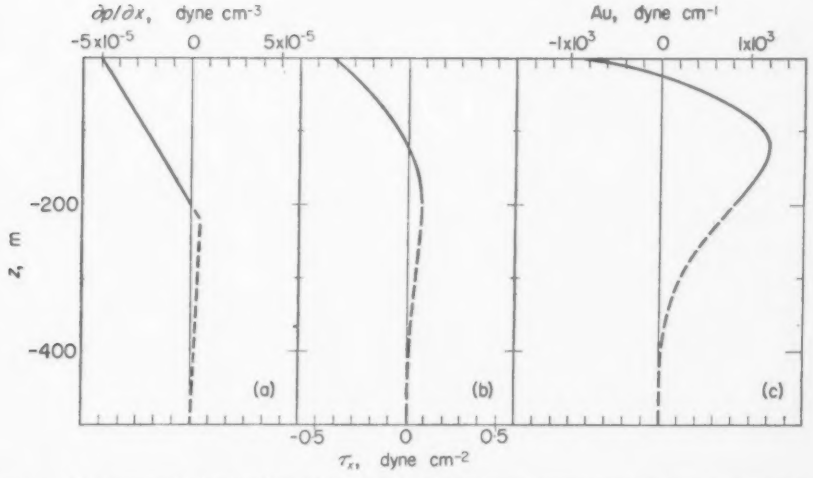

FIG. 1. (a) Assumed distribution with depth of east-west component of pressure gradient. (b) Resulting distribution of stress. (c) Resulting distribution of product of zonal velocity and coefficient of vertical eddy viscosity.

Integration of the x-component equation along the vertical from level z to the surface gives

$$\tau_x = (\tau_x)_0 - \int_z^0 \frac{\partial p}{\partial x}\, dz. \tag{7}$$

A distribution of $\tau_x(z)$ is derived from (7) for a given vertical distribution of $\partial p/\partial x$ (Fig. 1a) and an assumed value of the surface stress. The result is to be regarded as illustrative of a possibility derivable from a longitudinal component of the pressure

gradient, the feature which FOFONOFF and MONTGOMERY (1955) have shown to be essential for an undercurrent. The sea surface slopes upward toward the west along the equator, so $\partial p/\partial x$, therefore, is negative at the surface. Isotherms in the upper part of the thermocline layer have the reverse slope, and the magnitude of the pressure gradient is reduced to less than 20 per cent of the value at the surface by a depth of some two hundred metres. The solid-line portion of the curve for $\partial p/\partial x$ (Fig. 1a) is representative of the pressure gradient along the equator in the eastern Pacific.

The stress τ_x is seen (Fig. 1b) to increase with depth from the assumed negative value at the surface to a zero value just below a depth of 100 m. Since the stress then becomes positive, the depth at which τ_x becomes zero is of particular significance. The vertical shear in the zonal motion vanishes, and the velocity component must exhibit an extreme value. This depth of zero stress is obviously dependent upon the distribution of $\partial p/\partial x$ and the value of the surface stress.

The conclusion about an extreme value of the velocity component existing at the level of zero stress follows from the expression of shearing stress in the form

$$\tau_x = A \frac{\partial u}{\partial z}, \tag{8}$$

where A is the vertical coefficient of eddy viscosity. The value of A is known only within wide limits, e.g. $1 < A < 300$ in g cm^{-1} sec^{-1}, and it is not independent of z. In spite of the uncertainties, it will be useful to see whether an undercurrent can in fact exist for a uniform, positive value of A. The results may be regarded as representative of an appropriate average value of A. Integration of (8) along the vertical gives

$$u = u_0 - \frac{1}{A}\int_z^0 \tau_x \, dz = u_0 + \frac{1}{A}\left(z\tau_x + \int_z^0 z \frac{\partial p}{\partial x} \, dz\right). \tag{9}$$

The right-hand member is obtained after integration by parts and substitution from (5). If velocity and stress components are negligible at some depth $z = -B$, then

$$u_0 = \frac{1}{A}\int_{-B}^0 \tau_x \, dz = -\frac{1}{A}\int_{-B}^0 z \frac{\partial p}{\partial x} \, dz, \tag{10}$$

which shows how the surface velocity is determined by a bottom boundary condition. Alternatively, a surface boundary condition is used, i.e., u_0 is given in (9).

The distribution of Au with z is determined from (9) for the previous example with the added assumption of a negative value of Au at the surface. Appropriate values for A would lie between 10 and 50. The result (Fig. 1c) shows that an undercurrent is possible with these assumptions.

It is desirable to consider whether a bottom boundary condition can be satisfied, and the dashed portions of the curves in Fig. 1 relate to this inquiry. It has been assumed that at some depth, e.g., 500 m, the quantities $\partial p/\partial x$, τ_x, $\partial \tau_x/\partial z$, u and $\partial u/\partial z$ all become zero. This condition has been considered appropriate for an internal boundary within the fluid because it permits an underlying layer with no pressure gradient and no stress as introduced by SVERDRUP (1947), MUNK (1950), and others. Equations (7) and (9) show that for τ_x and u to reach zero at some depth $z = -B$, there must be a region of positive values of $\partial p/\partial x$ above the boundary as shown in Fig. 1a. It must, however, be pointed out that in the region where $\partial p/\partial x$ has relatively

Vol
6
1959

small numerical values, say $< 0{\cdot}5 \times 10^{-5}$ dyne cm^{-3}, the terms which have been neglected in the equation of motion may become important. Different boundary conditions might also be assumed, as, for example, those appropriate for a rigid boundary, where τ_x could have a small positive value at the bottom. Under such a condition, $\partial p/\partial x$ need not have any positive values ; but if $\partial p/\partial x < 0$ everywhere, then the velocity profile will have negative curvature everywhere between surface and bottom as it does in the solution for the special case of laminar flow in a long channel with wind directed along the axis.

The accuracy in the determination of $\partial p/\partial x$ is not sufficient to guarantee the existence of a region of small positive values of pressure gradient. Pressure is, of course, determined from observations of density by application of the hydrostatic equation. A reference level below 1000 m is desirable in the Pacific. ECKART (1958) has reviewed and analyzed empirical data on the equation of state of pure water and sea water. His results indicate an uncertainty of at least $\pm$ 20 dynamic cm in the calculation of dynamic height over a pressure interval of 1000 db as a result of uncertainties in the equation of state. However, REID (1959) has shown that the effect on calculation of horizontal pressure gradients at the surface may generally amount to an error of less than 5 per cent. Uncertainties in the horizontal temperature and salinity gradients introduce greater errors. Below a depth of 1000 m, the percentage error in the calculation of horizontal pressure gradient may increase to 50 per cent or more as a result of uncertainty in the value of the coefficient of thermal expansion. The distribution of $\partial p/\partial x$ assumed in the upper 200 m is comparable with that used by MONTGOMERY and PALMÉN (1940) from *Dana* stations along the equator in the Pacific. HIDAKA and NAGATA (1958) arrive at about the same distribution from more recent observations ; *Albatross* stations 92 and 111 (BRUNEAU, *et al.*, 1953) to 2000 m give values which are numerically somewhat larger ; AUSTIN (1958) obtains a mean value for $\partial p/\partial x$ of -5×10^{-5} dyne cm^{-3} at the surface with a reduction to less than 5 per cent of this value by a depth of 200 to 300 m. The first three sources do show a reversal in sign of $\partial p/\partial x$ somewhere below 200 m, but no conclusions can be drawn because of the uncertainties.

Wind stress on the sea surface is not precisely known, and an estimate is now made of the influence of this uncertainty. Equation (7) and the distribution of $\partial p/\partial x$ from Fig. 1a are used to calculate the depth $z = z_m$ at which τ_x becomes zero for given values of $(\tau_x)_0$. Values of $A\,(u_m - u_0)$ at $z = z_m$ are calculated from (9) in the form

$$A\,(u_m - u_0) = \int_{z_m}^{0} z\,\frac{\partial p}{\partial x}\,dz. \tag{11}$$

Results are shown (Table 1) for a range of values of surface stress.

Table 1. *Influence of wind stress*

$(\tau_x)_0$ (dyne cm^{-2})	z_m (m)	$A\,(u_m - u_0)$ (gm sec^{-2})
$-0{\cdot}3$	-75	$1{\cdot}1 \times 10^3$
$-0{\cdot}4$	-110	$1{\cdot}9 \times 10^3$
$-0{\cdot}5$	-200	$3{\cdot}3 \times 10^3$

According to HIDAKA's tabulation (1958), the average value of wind stress on the equator (100 °W–135 °W) in summer and fall is about $-0{\cdot}5$ dyne cm^{-2}. For a region farther west HIDAKA and NAGATA (1958, Fig. 1) use an annual mean value of $-0{\cdot}3$. REID (1948a) finds a value near $-0{\cdot}4$ for October–November in a region which extends from near the coast to about 150 °W. A value between $-0{\cdot}3$ and $-0{\cdot}4$ would lead to the best agreement between z_m (Table 1) and the observed depths of the core of the Undercurrent. Observed values of the difference $(u_m - u_0)$ would give a value of A of about 10 g cm^{-1} sec^{-1}.

Analysis of the distribution of τ_y and v with z at the equator would proceed in similar fashion. Observations indicate that the current is mainly zonal which means that τ_y and, therefore, $\partial p/\partial y$ should be relatively small or zero at the equator. This conclusion is supported to a degree by observations since isotherms in meridional sections tend to show a maximum or minimum depth at or near the equator.

If (2) is differentiated with respect to y, after neglect of acceleration and lateral friction the result for $\phi = 0°$ is

$$\beta\rho u = -\frac{\partial^2 p}{\partial y^2} + \frac{\partial^2 \tau_y}{\partial y\, \partial z}. \tag{12}$$

where $\beta = \partial f/\partial y$. If τ_y is sufficiently small not only at the equator but near the equator, then the last term in (12) may be neglected and the further approximation is

$$\beta\rho u \doteq -\frac{\partial^2 p}{\partial y^2}. \tag{13}$$

APPLICABILITY OF APPROXIMATIONS

The magnitudes of terms neglected in equations (1)–(3) are now considered at the equator. Observations indicate that $\partial u/\partial y$ is zero or very small in the westward flow near the surface at the equator. The Undercurrent appears to be symmetrical about the equator and reasonably steady. For the motion in the meridional plane, the model of CROMWELL (1953) is used, and v is taken as zero at the equator and w positive and symmetrical about the equator. Estimates are made using the following characteristic velocity and distance components:

$$U = 10^2 \text{ cm sec}^{-1} \qquad L_x = 5 \times 10^8 \text{ cm}$$

$$V = 4 \text{ cm sec}^{-1} \qquad L_y = 10^7 \text{ cm}$$

$$W = 10^{-3} \text{ cm sec}^{-1} \qquad L_z = 10^4 \text{ cm}$$

The value chosen for U is appropriate for the velocity near the core of the Undercurrent. FOFONOFF and MONTGOMERY (1955) have suggested the value 4 cm sec^{-1} for the velocity of flow toward the equator at the depth of the Undercurrent. The value of W is consistent from continuity considerations with the value of V and is a value which has been suggested as appropriate for upwelling at the equator (YOSHIDA, *et al.*, 1953). Distance L_x corresponds to about 50 degrees of longitude along the equator, L_y to about 1 degree of latitude, and L_z to the depth of the core of the Undercurrent.

In the x-component equation (1), the neglected acceleration term $v(\partial u/\partial y)$ is certainly negligible at the equator if v is zero and u is symmetrical. However, the

terms $u\,(\partial u/\partial x)$ and $w\,(\partial u/\partial z)$ may have values of the order of 10^{-5} cm sec^{-2} which is comparable with the longitudinal component of the pressure gradient term in the Undercurrent. The term $(u\,\partial u/\partial x)$ is zero at the core and in the westward surface current which shows relatively constant velocities along the equator. This term is positive in the Undercurrent above the core and negative below because of the rise in the core from west to east. The term $w\,(\partial u/\partial z)$ is zero at the core and at the surface. If w is positive, this term has negative values above the core and positive values below. While these two acceleration terms may not be separately negligible between core and surface, they have a tendency to offset each other. Complete cancellation occurs if the water particles move *without acceleration* along streamlines which are inclined to the horizontal. If the above characteristic velocities W and U are used, the value of the slope of the streamlines is 50 m in 5000 km which is almost identical with the observed rise in the core from west to east.

Actually, the acceleration of the water particles may not be zero at the equator, and an estimate is now made of the separate influence of an acceleration which results entirely from a term of the form $w\,(\partial u/\partial z)$. Calculations are made under the assumption that the right-hand side of (5) equals $w\,(\partial u/\partial z)$. The stress is taken to be zero at $z_m = -100$ m and the distribution of $\partial p/\partial x$ in Fig. 1 is used. Vertical velocity is assumed to decrease linearly from a value of w_m at z_m to a value of zero at the surface. Resulting values of $(\tau_x)_0$ and $A\,(u_m - u_0)$ are given in Table 2 for two values of $\rho w_m/A$ along with comparative values for zero vertical velocity.

Table 2. Influence of vertical velocity

$\rho w_m/A$ (cm^{-1})	$(\tau_x)_0$ (dyne cm^{-2})	$A\,(u_m - u_0)$ (gm sec^{-2})
0	$-0{\cdot}38$	$1{\cdot}7 \times 10^{3}$
1×10^{-4}	$-0{\cdot}43$	$2{\cdot}0 \times 10^{3}$
$1{\cdot}6 \times 10^{-4}$	$-0{\cdot}48$	$2{\cdot}2 \times 10^{3}$

Appropriate values of A would lie in the range of 10 to 20 gm cm^{-1} sec^{-1}, and w_m would then be of the order 10^{-3} cm sec^{-1}. Observed values of $(u_m - u_0)$ and $(\tau_x)_0$ might just be sufficiently accurate to verify the importance of this acceleration term $w\,(\partial u/\partial z)$. Values are changed by about 25 per cent.

The neglected Coriolis term, $2\Omega w \cos\phi$, in equation (1) has a magnitude of about 10^{-7} cm sec^{-2} and is negligible. Any estimates of lateral friction terms are difficult to arrive at because of the arbitrariness in values of lateral eddy viscosity. If these terms are replaced by the single term ku (YOSHIDA, *et al.*, 1953 ; WYRTKI, 1956), then the equation for unaccelerated motion at the equator is

$$A\frac{\partial^2 u}{\partial z^2} = \frac{\partial p}{\partial x} + ku, \tag{14}$$

where A has been assumed constant. The left-hand member is negative at the core and $\partial p/\partial x$ is about -2×10^{-5} dyne cm^{-3} and u is about 10^{2} cm sec^{-1}. Therefore the maximum value of k is 2×10^{-7} g cm^{-3} sec^{-1}. YOSHIDA, *et al.*, and WYRTKI have used a value of k equal to 5×10^{-6} and A equal to 10^{2} g cm^{-1} sec^{-1}. Equation

(14) may be solved with these values and a distribution of $\partial p/\partial x$ given in Fig. 1. For $(\tau_x)_0$ equal to $-0{\cdot}4$ dyne cm^{-2} and u_0 equal to -25 cm sec^{-1}, the value of u increases to -20 at 30 m depth and then decreases. Results bear little resemblance to observations and indicate that lateral friction has been overemphasized.

In the y-component equation (2), the neglected acceleration terms are all zero at the equator if v is zero. Terms neglected in the further approximation (13) are of the order of magnitude 10^{-13} g cm^{-3} sec^{-2}, which is to be compared to 10^{-11}, the order of magnitude of $\beta\rho u$.

In the z-component equation (3), the neglected acceleration terms and the Coriolis term $2\Omega u \cos\phi$ are all zero at the equator or are orders of magnitude smaller than ρg. The derivatives with respect to x and y of the neglected terms are also orders of magnitude smaller than the derivatives of terms retained.

COMPARISON WITH OTHER CALCULATIONS

SVERDRUP (1947) and REID (1948a) introduced the net mass transport from surface to great depth as a variable in place of velocity. Their theory differs from the calculation methods in that the distribution of density is not assumed, but a comparison with observations of the Undercurrent is nevertheless pertinent. Equation (4) and the previously indicated approximations to equations (1) and (2) were integrated along the vertical, and integrated mass transport and integrated pressure were calculated from values of wind stress for the eastern equatorial Pacific. Calculation of the zonal component of mass transport, M_x, requires a boundary condition which was satisfied by taking M_x equal to zero at the coastline on the east. The values of M_x at the equator show a net transport to the west away from the eastern coastline. The question is whether this result implies no undercurrent, and the answer lies in the interpretation of the condition $M_x = 0$. SVERDRUP and REID assume $(\tau_x)_0 \neq 0$ at the eastern boundary, and this means that $\partial p/\partial x \neq 0$ at this boundary at the equator. As a consequence of (1), neither τ_x nor u is zero at all depths, and the boundary is *not* to be interpreted as a vertical wall. Under these circumstances imposition of the condition $M_x = 0$ is somewhat arbitrary and must be interpreted to apply at some distance from the actual coastline with an intervening boundary layer in which terms omitted in equations (1) and (2) are important. An undercurrent may be present at the boundary as long as the *net* transport is zero. The results of SVERDRUP and REID need only imply that the transport of the undercurrent decreases westward from the boundary or that westward flow above or below the undercurrent increases.

WEENINK and GROEN (1952) solved similar approximations to (1), (2) and (3) for the surface velocity at the equator in terms of wind stress. Vanishing velocities at great depth are assumed, and REID's (1948b) model of the density distribution is applied. REID has pointed out certain deficiencies in his model near the equator. YOSHIDA (1955) has examined the results of WEENINK and GROEN and has shown how to remove a discontinuity in the meridional transport and a downward vertical velocity at the surface by application of a more complete density model. As pointed out by YOSHIDA, the method amounts to application of (10) to any vertical distribution of the pressure gradient obtained from observations or a model. Extension of the method to obtain velocities below the surface has been shown in the present paper to lead to an undercurrent.

Vol 6 1959

JERLOV (1953) applied approximation (13) to calculate the zonal component of velocity at the equator. REID (1948a) derived an expression similar to (12) but for integrated mass transport rather than velocity. TSUCHIYA (1955b) concludes that the approximations involved in these expressions are justified. According to (13), the sign of $\partial^2 p/\partial y^2$ should be negative at the depth of the Undercurrent. A trough in the isotherms or isolines of density favours such a sign and, as pointed out by WOOSTER and JENNINGS (1955), is probably related to the Undercurrent. Such a trough at or near the equator occurs in a number of meridional sections including those of KNAUSS (1960) and MONTGOMERY and STROUP (1960).

YOSHIDA *et al.*, (1953) replaced lateral friction terms by the terms $-ku$ and $-kv$ in the right-hand members of (1) and (2), respectively. Accelerations were neglected, and equation (4) was used to determine vertical velocity. Density observations near the equator were introduced into (3) to determine the pressure field. Calculations of u and v from (1) and (2) in terms of stress at top and bottom of the surface layer show no undercurrent. As pointed out previously the reason seems to be the over-emphasis on lateral friction through excessively large values of coefficient k. WYRTKI (1956) used essentially the same method to calculate mass transport of the surface layer.

HIDAKA and NAGATA (1958) calculated u and v from (1) and (2) with accelerations neglected. Wind stress and pressure gradients were assumed from observations. The results show an undercurrent, but the method cannot be considered valid because of the magnitude of field accelerations. If the results are used to calculate $v\,(\partial u/\partial y)$ at the equator, the order of magnitude of this neglected term is 10^{-4} cm sec^{-2} which is as large as the magnitudes of terms retained in (1). It should also be noted that the resulting undercurrent is not symetrical with the equator.

FOFONOFF and MONTGOMERY (1955) find the Undercurrent to be consistent with conservation of absolute vorticity in an equatorward flow of a layer of constant thickness. The model assumes a meridional circulation in which the water in the lower part of the surface layer flows equatorward and then moves upward at the equator. The acceleration term $v\,(\partial u/\partial y)$ is found to be important, but the treatment considers the conditions near but not right on the equator. Since this term does not enter the y-component equation of motion, there is no conflict with approximation (13).

SUMMARY AND CONCLUSIONS

1. The assumption is made that two important terms in the x-component (east-west) equation of motion at the equator are the pressure gradient and the frictional stress across horizontal planes. The two terms are equated, and the profile of zonal velocity is calculated from values of surface velocity, wind stress, and horizontal pressure gradient for a constant vertical eddy coefficient. The results give an undercurrent with certain features such as the reversal in direction of current at shallow depths which compare favourably with observations. The calculated depth of the core of maximum eastward velocity is independent of the particular value of the eddy coefficient. The rise of the core toward the east suggests a relatively reduced surface wind stress or an increased pressure gradient.

2. The motion appears to be steady, but two of the field acceleration terms, $u\,(\partial u/\partial x)$ and $w\,(\partial u/\partial z)$, are probably not negligible. FOFONOFF and MONTGOMERY

(1955) have demonstrated the importance of the term $v(\partial u/\partial y)$ in the region where water moves toward the equator. This term may be neglected at the equator. However, the implication is that acceleration terms may not be neglected for the determination of currents *both on and near the equator*. This conclusion is important because theories of the surface circulation which include the equatorial regions have neglected acceleration. Conversely, friction is important in the core of the Undercurrent ; and absolute vorticity may not be conserved in the immediate vicinity of the equator. Velocity profiles observed by KNAUSS (1960) would suggest that friction in the core is important to at least 0·5 degrees of latitude north and south of the equator if the value of A appropriate at the equator is used.

3. JERLOV (1953) and others have differentiated the geostrophic formula for zonal motion to remove the indeterminancy at the equator. Indications are that the approximation may have validity, but its usefulness is limited by the difficulty in obtaining sufficiently accurate estimates of $\partial^2 p/\partial y^2$. Nevertheless, this approximation seems to offer the most hope for a successful formula for calculation of currents at the equator, because acceleration and friction terms appear to be negligible in this particular equation.

4. Other formulae include lateral friction. Results are unrealistic because of excessively large friction coefficients or are uncertain because of large accelerations.

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Vol
6
1959

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Vol.
6
959-60

Wind-drift near the Equator

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Abstract—The principal result of the model is to show that a linear theory constructed from Ekman spiral and geostrophic current does not suffer singularities at the equator and even exhibits some qualitative similarity to the Equatorial Undercurrent, but that in order to account for the observed amplitude of the phenomenon, non-linearities are not altogether negligible.

(1) INTRODUCTION

IN EKMAN's well-known solution of drift of water in a rotating homogeneous ocean of infinite depth when acted upon by a steady stress applied at the surface, the depth of the spiral, and amplitude of the currents increase without limit as the vertical component of rotation (or latitude) approaches zero. It is possible to add an arbitrary infinitesimal deformation of the free surface to Ekman's homogeneous model and an associated pattern of geostrophic flow – independent of depth – whose Coriolis force just balances the arbitrary pressure gradient. In fact, Ekman uses such additional geostrophic velocities to satisfy conditions of no flow through coastal barriers. The one-dimensional nature of the simplified equations used by Ekman does not permit satisfying the true boundary condition at a vertical coastal barrier : namely, that the velocity normal to the barrier shall vanish at all depths. Instead of trying to solve the details of this more complicated problem, which evidently would involve a higher-order set of dynamical equations, Ekman simply adds a pressure field sufficient to cause the total vertically integrated component of velocity normal to the coast to vanish. When the ocean is of finite depth, a finite geostrophic current is required, and when the bottom is rigid it is necessary to allow for a second frictional boundary layer, or current spiral at the bottom. This latter complication can be avoided by asserting that the bottom boundary condition is one of no stress – such as can be achieved by letting the viscosity become very small near the bottom.

In these ways Ekman is able to avoid the indeterminacy of additive geostrophic motion in his simple one-dimensional (vertical) dynamical system, to make an approximate allowance for the influence of lateral boundaries, and also to compute patterns of flow in homogeneous ocean models under various distributions of wind and coasts.

The purpose of the present paper is to show that this simple type of analysis can be extended across the equator, and that there is indeed no singularity at the equator. Thus, in principle, there is no reason to suppose that Ekman's solution ' blows up ' at the equator, as is sometimes asserted in qualitative discussions. However, it will be shown that the model is consistent only when the amplitude of the velocity is kept at an unrealistically low value by imposing a very small value of wind stress, or a large eddy viscosity. When realistic values are introduced and the solution for the velocity field is used to compute the magnitude of the neglected inertia terms it is found that they are not truly negligible and that the model presented here becomes

Vol
6
1959

inconsistent. Therefore it seems that the model does not apply as it stands to the Pacific Equatorial Undercurrent (Cromwell Current). From a pedagogical point of view however the model may be of some value because it exhibits qualitative features similar to the Cromwell Current, because in an ocean where mean wind stress along the equator is less (Indian Ocean) it might actually apply, and because it embraces the very simplest linear type of dynamical scheme which one is bound to investigate before pursuing more complex models including lateral friction, inertial terms, etc.

At one time Dr. VERONIS and I considered introducing the heat-transport equation in a linearized form such as used by LINEYKIN (1955, 1957) and STOMMEL and VERONIS (1957) and developing the temperature, pressure and velocity fields about the equator in a power series in latitude (y), but such an analysis is also inapplicable to the actual Cromwell Current because of the neglect of inertial terms, and we decided that it did not reveal enough additional information to publish. In this same issue Dr. CHARNEY shows how the inertial terms can be included in the simple homogeneous model, and in the paper after that Dr. VERONIS shows how the thermal field can also be included.

(2) THE SOLUTION OF THE LINEAR PROBLEM

The model consists of a homogeneous layer of water of uniform depth h straddling the equator. Acting upon the upper surface there is a westward wind stress τ. It does not appear likely that a current like the Cromwell Current is caused by a detailed structure of the wind stress as a function of latitude, so the wind stress is assumed to be uniform. The x-axis is direct toward the east, the y-axis northward, the z-axis upward and $z = 0$ at the bottom of the layer. The eddy viscosity is a constant A.

The linearized equations of motion, in which latitude enters only parametrically through the vertical component of the earth's angular velocity ω are of the form :

$$-2\omega v = A\frac{\partial^2 u}{\partial z^2} - \frac{1}{\rho}\frac{\partial p}{\partial x}$$

$$2\omega u = A\frac{\partial^2 v}{\partial z^2} - \frac{1}{\rho}\frac{\partial p}{\partial y}$$

where p is pressure, ρ density, and u, v the x and y components of velocity. We assume the water to be homogeneous in density so that $\partial p/\partial x$ and $\partial p/\partial y$ are independent of depth. The solutions are of the form

$$u + iv = \sum_1^4 c_i e^{\alpha_i z} + G_1 + i G_2$$

where

$$\alpha_1 = r(1+i), \quad \alpha_2 = r(1-i), \quad \alpha_3 = r(-1+i), \quad \alpha_4 = r(-1-i)$$

$$r = \sqrt{\omega/A}$$

G_1 and G_2 are real constants. We take $z = 0$ at the bottom where we assume $A\frac{\partial u}{\partial z} = A\frac{\partial v}{\partial z} = 0$. Making the bottom a plane of no stress has the virtue of avoiding the necessity of calculating an additional functional layer at the bottom, thus simplifying the analysis.

At the top surface, $z = h$, the stress τ is taken to be only in the x-direction. These boundary conditions are enough to fix the four constants c_i. The additive constants G_1, G_2 corresponding to geostrophic flow, are evaluated by taking the vertically integrated transports equal to zero. This might at first seem to be a very arbitrary step. It can be perhaps justified by stating first that since we will eventually allow ω to vary parametrically ($\omega = \beta y$) the well-known 'curl' equation

$$\beta \int_0^h v dz = \text{curl}\,(\tau/\rho)$$

hence, since curl $(\tau/\rho) = 0$, the vertical integral of v must also vanish everywhere. The continuity equation

$$\frac{\partial}{\partial x}\int_0^h u dz + \frac{\partial}{\partial y}\int_0^h v dz = 0$$

reduces to the form

$$\frac{\partial}{\partial x}\int_0^h u dz = 0$$

If, therefore, $\int_0^h u dz$ is set equal to zero at an eastern meridional boundary it also must vanish everywhere. The solution of the problem can now be written in the form :

$$u + iv = \frac{\tau}{\mu\alpha_1 \rho}\left(\frac{\cosh \alpha_1 z}{\sinh \alpha_1 h} - \frac{1}{\alpha_1 h}\right)$$

In order to evaluate the velocity components u and v as functions of z and ω it is necessary to separate real and imaginary parts of the right hand side. Writing $Z = rz$ and $H = rh$ we obtain the following expressions :

$$\frac{u}{\left(\frac{\tau h}{2A\rho}\right)} = \frac{1}{H(\sinh^2 H + \sin^2 H)}\{\cosh Z \cos Z \sinh H \cos H + \sinh Z \sin Z \cosh H \sin H + \sinh Z \sin Z \sinh H \cos H - \cosh Z \cos Z \cosh H \sin H\}$$

$$\frac{v}{\left(\frac{\tau h}{2A\rho}\right)} = \frac{1}{H^2} + \frac{1}{H(\sinh^2 H + \sin^2 H)}\{\sinh Z \sin Z \sinh H \cos H - \cosh Z \cos Z \cosh H \sin H - \cosh Z \cos Z \sinh H \cos H - \sinh Z \sin Z \cosh H \sin H\}$$

We can regard the H as containing the essential parametric variation of the quantities with latitude. The following table gives some computed values

Vol 6 1959

z/h			H = 0	$\pi/2$	π	$\to \infty$
	1·0	$\frac{2A\rho}{\tau h} \cdot u$	0·666	0·584	0·285	$1/H$
		$\frac{2A\rho}{\tau h} \cdot v$	0·000	− 0·180	− 0·186	$-1/H$
	0·5	$\frac{2A\rho}{\tau h} \cdot u$	− 0·083	− 0·081	− 0·063	0
		$\frac{2A\rho}{\tau h} \cdot v$	0·000	0·013	0·038	$1/H^2$
	0·0	$\frac{2A\rho}{\tau h} \cdot u$	− 0·333	− 0·254	− 0·027	0
		$\frac{2A\rho}{\tau h} \cdot v$	0·000	0·151	0·128	$1/H^2$

At $H = 0$, which corresponds to the equator, the expressions can be reduced to the following form,

$$u + iv = \frac{\tau h}{2A\rho}\left(\frac{z^2}{h^2} - \frac{1}{3}\right) + i\,0$$

a simple parabolic velocity profile, with no north-south component. The case $H \to \infty$ corresponds to the combination of a thin surface Ekman layer underlain by an equatorward meridional geostrophic component. The figure shows the velocity vectors on either side of the equator. Quantitatively, the mean stress τ at the equator is about − 0·2 dynes cm^{-2} in the Pacific, and the layer depth (core of undercurrent) is $h = 10^4$ cm. The observed amplitude of the undercurrent at $H = 0$ and $z/h = 0$, is about 150 cm sec^{-1}, so that eddy viscosity A is 2·2 cm^2 sec^{-1}. According to the model the undercurrent vanishes at a latitude where $H = \pi$, so that using these values the northern border of the undercurrent is at a latitude corresponding to $\omega = A \times 10^{-7}$ sec^{-1}, or about 0·15°N. The observed boundary is more nearly at 1·5° latitude, so this agreement is not good. If the eddy viscosity is increased to 22·0 cm^2 sec^{-1}, the computed northern boundary of the undercurrent agrees with that observed, but the velocities in the core of the undercurrent are reduced to 15 cm sec^{-1}, which is much too low. Therefore, it appears that this linearized model does *not* apply to the Pacific Equatorial Undercurrent. Furthermore, using the above values of the wind stress and the two choices of eddy viscosity, a calculation shows that the non-linear inertial terms are of the same order of magnitude as the other terms in the neighbourhood of the equator, and should not be neglected.

A reviewer has indicated several obscurities in the presentation which need to be cleared up. First, strictly speaking, the homogeneous layer lies upon a deeper resting layer of greater density. If there are pressure gradients in the homogeneous upper layer, but none in the lower layer, there must be a slope in the interface between them.

Therefore the upper layer cannot be of precisely uniform depth. Also it is worthwhile to emphasize that there is a local balance of pressure-gradient force on the layer and wind stress, and the uniformity of one guarantees that of the other. And finally,

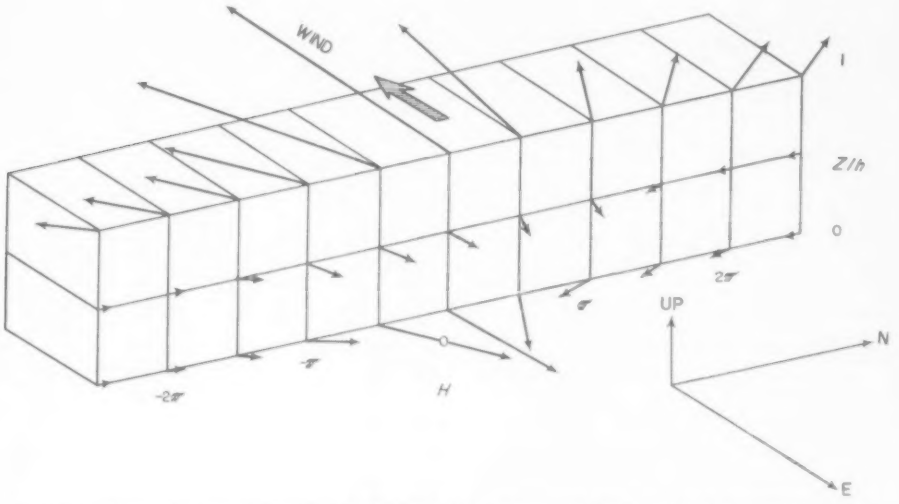

FIG. 1. Schematic three-dimensional diagram of the combined frictional and geostrophic flow field in the neighbourhood of the equator, not showing the vertical component of velocity (which can in principle be determined from continuity).

the current transport field obtained in this model is very similar to that pictured by CROMWELL (1953, Fig. 6), and we might regard the present model as a mathematical analysis of Cromwell's dynamical description of the currents produced by an east wind at the equator.

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Vol
6
1959

Non-linear theory of a wind-driven homogeneous layer near the Equator

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Abstract—This article deals with the equatorial motion produced by a uniform east wind blowing over a homogeneous ocean of constant depth. It is shown by a dimensional argument that for the range of parameters applicable to the Cromwell Current, the inertial forces are equally important with the frictional forces. The motion is calculated for several values of the coefficient of eddy-viscosity. The calculations give an undercurrent whose intensity increases with decreasing values of the eddy viscosity and which, for sufficiently small values, resembles the Cromwell Current in several of its principal features, in intensity, in width, in horizontal and vertical velocity variation with depth.

An analogy is drawn between the western boundary currents and the equatorial undercurrents. It is shown that both are in a sense, inertial boundary currents, possessing frictional boundary-layers near the boundaries, the western coasts in the former instance and the equator in the latter.

Vol.
6
959-60

(1) INTRODUCTION

THE Cromwell Current flows along the equator and its absolute angular momentum exceeds that of the surrounding waters. To maintain this angular momentum against counteracting frictional torques a west-east pressure gradient would seem to be required. At the equator such a pressure gradient cannot be balanced by a horizontal Coriolis force component but must be balanced by frictional or inertial forces*. The singular character of the current can evidently be attributed to this circumstance. In the present note we investigate the balance of forces near the equator and calculate the equatorial undercurrent for a simple hydrodynamical analogue of the Cromwell Current.

The remarkable symmetry and uniformity of this current suggest that its essential structure is independent of the details of the surface wind stress. We therefore postulate : (1) that this stress is uniform and westward, (2) that the flow is symmetric about the equator, and (3) that it is either independent of longitude or is at most a slowly varying function of longitude. For simplicity we also assume the model ocean to be homogeneous and of a uniform depth H in the undisturbed state. To make possible a comparison of theory with observation we then identify this depth with that of the centre of the equatorial thermocline. This is done on the assumption that the high gravitational stability of the thermocline inhibits the transmission of pressure forces and momenta to such an extent that the thermocline may be approximated by a rigid boundary.

The frictional forces are taken into account by postulating a constant value for the kinematic coefficient of eddy viscosity, ν. Since in reality the eddy coefficient is not constant and its values are not reliably known, the calculations will merely show

*The horizontal Coriolis force component due to vertical motion fails by more than an order of magnitude to balance the observed pressure gradients (see ARTHUR, this journal).

that qualitative agreement with observation is obtained for a value of ν close to 15 $cm^2\,sec^{-1}$. This value may therefore be taken as a kind of mean value within the Cromwell Current. The theory to be presented, however, is logically independent of the assumptions required for application to the real ocean and would apply just as well to a laboratory model experiment in which ν is the constant value of the molecular viscosity.

(2) DETERMINATION OF THE EQUATORIAL PRESSURE GRADIENT

At some distance from the equator the inertial forces and the vertical velocities are small. Hence, in the usual notation, with x pointing eastward, y northward and z upward, the equations of motion

$$\left.\begin{aligned} \rho\,(du/dt + 2\Omega w\cos\phi - 2\Omega v\sin\phi) &= -p_x + (\tau^{(x)})_z, \\ \rho\,(dv/dt + 2\Omega u\sin\phi) &= -p_y + (\tau^{(x)})_z, \\ (\tau^{(x)} = \nu u_z, \quad \tau^{(y)} &= \nu v_z), \end{aligned}\right\} \tag{2.1}$$

reduce to

$$\left.\begin{aligned} -\rho f v &= -p_x + (\tau^{(x)})_z, \\ \rho f u &= -p_y + (\tau^{(y)})_z. \end{aligned}\right\} \tag{2.2}$$

Near the equator the Coriolis parameter f is closely approximated by

$$f = 2\Omega y/R \equiv \beta y, \tag{2.3}$$

where R is the radius of the earth. Taking the curl of equations 2.2 and integrating with respect to z from the bottom ($z = 0$) to the top ($z = H$) on the layer, we get

$$\beta\rho\int_0^H v\,dz + f\rho\int_0^H (u_x + v_y)\,dz = (\tau^{(x)})_y - (\tau^{(y)})_x\Big|_{z=0}^{z=H} = 0, \tag{2.4}$$

since the surface stress is constant and we may assume that the bottom stress is negligible. Integration of the equation of continuity,

$$u_x + v_y + w_z = 0, \tag{2.5}$$

with respect to z shows that the divergence integral in equation 2.4 also vanishes, since w vanishes at $z = 0$ and is negligible at $z = H$. Thus we find that the vertically integrated meridional volume transport is zero. From this result it follows, by vertical integration of the first of equations 2.2, that

$$-Hp_x = -(\tau^{(x)})_{z=H} \equiv \tau. \tag{2.6}$$

Since the above equation holds up to the point at which the inertial forces become large, i.e. to within two or three degrees of the equator, it must hold also at the equator. This is because τ does not change, and the zonal pressure gradient cannot change appreciably without producing a very large zonal variation of the meridional pressure gradient and a correspondingly large zonal non-uniformity of the current, such non-uniformity being contrary to our initial assumption.

(3) THE BALANCE OF FORCES NEAR THE EQUATOR

In virtue of the foregoing analysis, the steady-state equations of motion and continuity in the vicinity of the equator may be written

$$\left.\begin{aligned} vu_y + wu_z - \beta yv &= P + \nu u_{zz}, \\ vv_y + wv_z + \beta yu &= Q + \nu v_{zz}, \end{aligned}\right\} \tag{3.1}$$

$$v_y + w_z = 0, \tag{3.2}$$

where P and Q are the pressure force components in the x and y directions respectively, and $P = \tau/\rho H$.

It will follow *a posteriori* that the ratio of the vertical to the horizontal scales of motion is so small, even at the equator, that the flow may be assumed to be hydrostatic, with P and Q independent of depth. Likewise we may justify *a posteriori* the omission of the vertical Coriolis force term from the first equation of motion.

On the assumption that the inertial terms predominate and that the velocity components are smooth functions of z on the interval $0 \leqslant z \leqslant H$, we find

$$\left.\begin{aligned} u, v &\sim (P^2/\beta)^{1/3}, \\ w &\sim H(P\beta)^{1/3}, \\ y &\sim (P/\beta^2)^{1/3}, \end{aligned}\right\} \tag{3.3}$$

and the equations of motion and continuity in the non-dimentional quantities $(U, V) = (P^2/\beta)^{-1/3}(u, v)$, $W = (H^3 P\beta)^{-1/3} w$, $\eta = (P/\beta^2)^{-1/3} y$, $\zeta = H^{-1} z$ become

$$\left.\begin{aligned} VU_\eta + WU_\zeta - \eta V &= 1 + cU_{\zeta\zeta}, \\ VV_\eta + WV_\zeta + \eta U &= Q' + cV_{\zeta\zeta}, \end{aligned}\right\} \tag{3.4}$$

$$V_\eta + W_\zeta = 0, \tag{3.5}$$

where

$$c = \nu/H^2(P\beta)^{1/3}, \quad Q' = Q(P\beta)^{-2/3}. \tag{3.6}$$

These equations are subject to the boundary conditions :

$$U_\zeta(\eta, 1) = -1/c, \quad V_\zeta(\eta, 1) = 0, \tag{3.7}$$

which follow from equation 2.6 and the condition that the y-component of the wind stress is zero; and

$$U(\eta, 0) = V(\eta, 0) = W(\eta, 0) = W(\eta, 1) = 0, \tag{3.8}$$

which follow from the vanishing of U, V and W at the bottom boundary and W at the free surface.

We take $H = 150$ m as the mean depth of the centre of the equatorial thermocline between 2 °S and 2 °N latitude. By equation 2.6 a wind stress of 0·45 dynes cm^{-2} gives $P = 0{\cdot}45/1{\cdot}5 \times 10^4 = 3 \times 10^{-5}$ cm sec^{-2}, corresponding to a rise of the free surface toward the west of 3 cm per 1000 km. This value may be compared with the value 4 cm per 1000 km given by MONTGOMERY and PALMÉN (1940).

If one takes ν to be in the range 10 to 100 cm^2 sec^{-1}, c is found to lie in the range 1/40 to 1/4. With reference to equations 3.4 it would then appear that our assumption of the predominance of the inertial terms is correct. Actually, however, it is found that the frictional terms are equally important, not merely at the upper and lower boundaries, but throughout the layer. This apparent contradiction can be traced to the fact that for small c (or ν) W fluctuates in sign, so that the inertial term WU_ζ in the first of equations 3.4 cannot balance the pressure force everywhere.

(4) SOLUTION AT THE EQUATOR

Setting $\eta = V = 0$ in the equations of motion, after first differentiating the second with respect to η and utilizing the continuity equation, we obtain

$$\left.\begin{aligned} WU_\zeta &= 1 + cU_{\zeta\zeta}, \\ W_\zeta{}^2 - WW_{\zeta\zeta} + U &= G - cW_{\zeta\zeta\zeta}, \end{aligned}\right\} \tag{4.1}$$

where $G = (Q_\eta')_{\eta=0}$. These equations are subject to the boundary conditions

$$\left.\begin{aligned} U(0) &= 0, \quad U_\zeta(1) = -1/c, \\ W(0) &= W_\zeta(0) = W(1) = W_{\zeta\zeta}(1) = 0, \end{aligned}\right\} \tag{4.2}$$

which follow from equations 3.7, 3.8 and 3.5. Because of their non-linearity we must resort to a numerical method for their solution.

While the steady-state equations are very difficult to solve, even numerically, it is quite simple to solve the time variable problem. If we define a non-dimensional time T by $T = t(P\beta)^{1/3}$, the time-variable equations become

$$U_T + WU_\zeta = 1 + cU_{\zeta\zeta}, \tag{4.3}$$

$$-W_{T\zeta} + W_\zeta{}^2 - WW_{\zeta\zeta} + U = G - cW_{\zeta\zeta\zeta}. \tag{4.4}$$

Integrating the second with respect to ζ and utilizing the conditions 4.2, we find

$$W_T - WW_\zeta = G(\zeta) - \zeta G(1) + cW_{\zeta\zeta}, \tag{4.5}$$

where

$$G(\zeta) = \int_0^\zeta (U - 2WW_{\zeta\zeta})\, d\zeta - cW_{\zeta\zeta}(0), \tag{4.6}$$

and

$$G(1) = G. \tag{4.7}$$

Equations 4.3 and 4.5 are a set of parabolic partial differential equations closely analogous to the conduction equation. Finite-difference methods of solution for this type of system are known (cf. RICHTMYER, 1957). We set the quantities U and W equal to zero initially and calculate their time evolutions until, because of dissipation, they approach a steady-state sufficiently closely. The following finite-difference scheme has been found to be stable and to give good results with an IBM 704 electronic computer: The interval $0 \leqslant \zeta \leqslant 1$ is divided into 20 equal parts at the points $j\,\Delta\zeta$ ($j = 0, 1, \ldots, 20$), and discrete times are defined by $T = n\,\Delta T$ ($n = 0, 1, 2, \ldots$). Denoting quantities at the points $j\,\Delta\zeta$ and the times $n\,\Delta T$ by the subscripts j and the superscripts n, we replace equations 4.3, 4.5 and 4.6 by their finite-difference analogues

$$\frac{U_j^{(n+1)} - U_j^{(n)}}{\Delta T} = -W_j^{(n)} \frac{\Delta_j U^{(n)}}{2\Delta\zeta} + c\frac{\Delta_j^2 U^{(n)}}{(\Delta\zeta)^2} + 1, \tag{4.8}$$

$$\frac{W_j^{(n+1)} - W_j^{(n)}}{\Delta T} = W_j^{(n)} \frac{\Delta_j U^{(n)}}{2\Delta\zeta} + c\frac{\Delta_j^2 U^{(n)}}{(\Delta\zeta)^2} + G^{(n)} - j\,\Delta\zeta\, G_{20}{}^{(n)}, \tag{4.9}$$

$$G_j^{(n)} = (\tfrac{1}{2}U_0 + U_1 + \ldots U_{j-1} + \tfrac{1}{2}U_j)^{(n)} - 2cW_1^{(n)}/(\Delta\zeta)^2 -$$

$$-2(W_1\Delta_1{}^2 W + W_2\Delta_2{}^2 W + \ldots + W_{j-1}\Delta^2{}_{j-1} W + \tfrac{1}{2}W_j\Delta_j{}^2 W)^{(n)}/\Delta\zeta, \tag{4.10}$$

Vol 6 1959

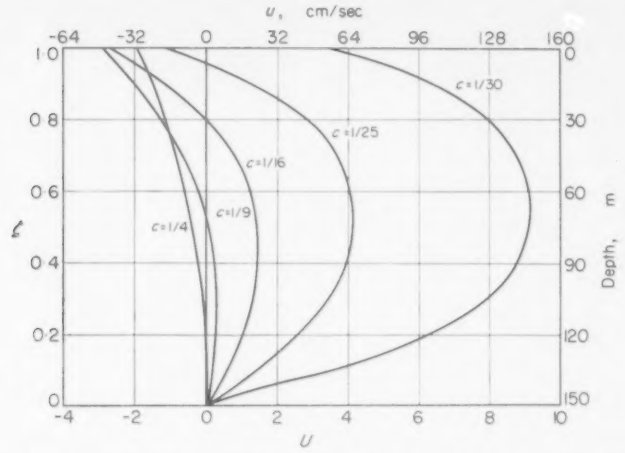

FIG. 1. West-east velocity at the equator as a function of depth for different values of the non-dimensional parameter $c = \nu/H^2(P\beta)^{1/3}$.

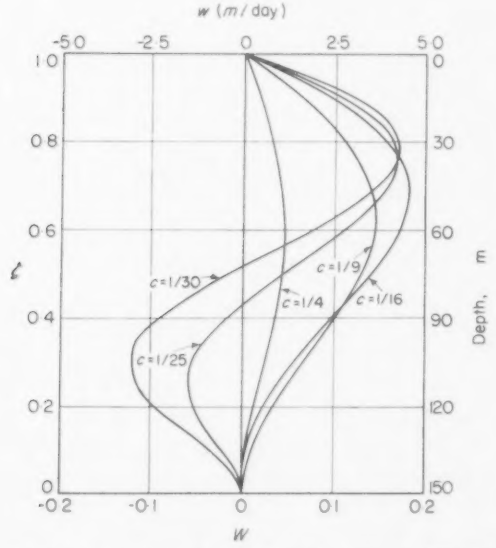

FIG. 2. Vertical velocity at the equator as a function of depth for different values of the non-dimensional parameter $c = \nu/H^2(P\beta)^{1/3}$.

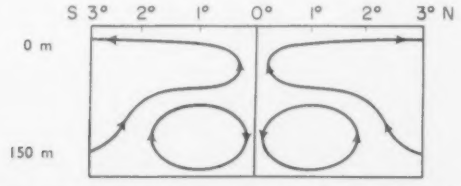

FIG. 3. Schematic representation of the vertical circulation in the vicinity of the equator.

in which $\Delta_j \alpha \equiv \alpha_{j+1} - \alpha_{j-1}$ and $\Delta_j^2 \alpha \equiv \alpha_{j+1} - 2\alpha_j + \alpha_{j-1}$. These equations define an iterative sequence for $U^{(n+1)}$ and $W^{(n+1)}$ in terms of $U^{(n)}$ and $W^{(n)}$. The time step ΔT was chosen so as to satisfy the computational stability criterion :

$$\frac{c\Delta T}{(\Delta\zeta)^2} \leqslant \tfrac{1}{2}. \tag{4.11}$$

In practice between 1000 and 2000 iterations are required for convergence to a steady state for both small and large values of c. The smaller time step needed for larger values of c was compensated for by the more rapid convergence rate due to the larger dissipation.

The calculations were performed for $c = 1/4$, $1/9$, $1/16$, $1/25$ and $1/30$, corresponding to $\nu = 107$, 47, 27, 17 and 14 cm^2 sec^{-1}. The calculated U and W profiles are shown in Figs. 1 and 2. The U-unit being $(P^2/\beta)^{1/3} \cong 16$ cm sec^{-1}, and the W-unit being $H(P\beta)^{1/3} \cong 2{\cdot}9 \times 10^{-2}$ cm sec$^{-1} \cong 25$ m per day.

It will be seen from Fig. 1 that the velocities become comparable to those in the Cromwell Current when c lies between $1/25$ and $1/30$ (ν between 17 and 14 cm^2 sec^{-1}). Smaller values of c give impossibly large velocities. Fig. 2 shows the vertical velocity at the equator. A noteworthy feature is the reversal of this velocity at about 80 m. This feature is in agreement with the evidence from the temperatures, salinities and oxygen contents presented by KNAUSS. It suggests a double-celled circulation with upwelling near the surface and descending motion below about 80 m. A schematic representation of this circulation, made to fit the supposed circulation away from the equator (SVERDRUP, 1946), is shown in Fig. 3. It is also noteworthy that the vertical velocities reach values of nearly 5 m per day. These are so large that it would seem not impossible to measure them by some direct means.

The decisive role played by the non-linear field accelerations is clearly brought out by comparison of the curves in Fig. 1 with the linear solution, $U = -\zeta^2/2c$, which gives a velocity directed towards the west.

(5) THE CROMWELL CURRENT AND THE GULF STREAM, AN ANALOGY

In the open oceans the large-scale flows produced by wind action or by density differences are very nearly quasi-geostrophic. The ratio of the horizontal components of relative acceleration and Coriolis force varies typically from 10^{-4} to 10^{-2}. Hence, as a rule, one may employ the geostrophic approximation in the mathematical analysis of these motions. Exceptions to this rule occur at rigid boundaries where the depressed orders of the geostrophic equations of motion make it impossible to satisfy the condition of vanishing velocity, and at the equator where the east-west Coriolis force component due to the horizontal motion vanishes. In such regions the question of the relative importance of frictional and inertial forces can be resolved by a kind of dimensional analysis familiar in boundary-layer theory (e.g., CARRIER, 1953). Similar reasoning has been employed by the author to establish the quasi-geostrophic equations for the atmosphere (1948) and for the oceans (1955a).

Let us first briefly consider the problem of the western boundary currents. These currents have been explained by MUNK and CARRIER (1950) as frictional boundary layers and by the author (1955b) as inertial boundary layers. Both theories are consistent. Which of them is the more applicable must depend on the magnitudes of the physical parameters involved. Let us consider a homogeneous ocean with a

Vol 6 1959

north-south boundary at $x = 0$ and, following MUNK and CARRIER, assume only a lateral eddy momentum exchange. We suppose that A, the lateral eddy coefficient of viscosity, and β, the y-derivative of the Coriolis parameter, are constant. Then the boundary-layer motion is governed by the vorticity equation.

$$\frac{\partial(\psi_{xx}, \psi)}{\partial(x, y)} + \beta\psi_x = A\psi_{xxxx} \tag{5.1}$$

where ψ is the stream-function for the horizontal velocity, and the vorticity $\nabla^2\psi$ is approximated by ψ_{xx}. This equation is subject to the conditions

$$\psi(0, y) = \psi_x(0, y) = 0, \quad \psi_y(\infty, y) = F(y). \tag{5.2}$$

Let L be a typical magnitude of y and U a typical magnitude of $F(y)$. If it is assumed that the inertial terms dominate, the characteristic magnitude of x, i.e. the width of the current, is found to be $\delta = (u/\beta)^{1/2}$, and the equation of motion in the non-dimensional quantities $\psi = L\,u\,\psi$, $x = \delta\xi$, $y = L\eta$, is

$$\frac{\partial(\Psi_{\xi\xi}, \Psi)}{\partial(\xi, \eta)} + \Psi_\xi = \frac{A}{u\delta}\,\Psi_{\xi\xi\xi\xi} \tag{5.3}$$

The relevant non-dimensional parameter is thus $\mu = A/u\delta = (\beta A^2/u^3)^{1/2}$. The value of u is of the order of 10 cm sec^{-1}, and STOMMEL (1955) finds A to be $9\ (+5) \times 10^5$ cm^2 sec^{-1} in the Gulf Stream. Taking the uppermost of his values for A we obtain 10^{-2} as the order of magnitude of the frictional term in the Gulf Stream. It would therefore appear that the inertial forces dominate the frictional forces throughout the main body of the stream. However, there must still exist a frictional sub-boundary layer next to the wall in order to reduce the current velocity to zero. Its dimensions can be ascertained by performing the stretching substitutions $\Psi = \mu^l x$, $\xi = \mu^m$. We find that the frictional term has the same order as the inertial and Coriolis terms when $l = 2/3$ and $m = 1/3$. Hence the width of the frictional sub-boundary layer is $\mu^{1/3}$, or about one-fourth to one-fifth the width of the Gulf Stream. This is in agreement with observation (WORTHINGTON, 1953; VON ARX, *et al.*, 1955).

To find the lateral extent of the Cromwell Current we adopt the approach of FOFONOFF and MONTGOMERY (1955) who have analysed this current with the aid of the barotropic vorticity equation. They assume that the flow is two dimensional and barotropic in some mean sense. It is then natural to postulate a constant lateral eddy viscosity as the dissipative factor. In this case the flow is governed by an equation exactly analogous to 5.1 except that in the expression for the vorticity x and y are interchanged. In place of equation 5.3 one obtains

$$\frac{\partial(\Psi_{\eta\eta}, \Psi)}{(\xi, \eta)} + \Psi_\xi = \frac{LA}{U\delta^2}\,\Psi_{\eta\eta\eta\eta} \tag{5.4}$$

where L is now a typical x-length, $\delta = (U/\beta)^{1/2}$ is a typical y-length, and U is a typical current velocity, i.e. we set $\psi = \delta\,U\,\Psi$, $x = L\xi$ and $y = \delta\eta$. We know that U is of the order 10^2 cm sec^{-1} and L is of the order 5000 km. Hence, taking Stommel's values for A we find the coefficient of the friction term to have the order 10^{-2}. The stretching substitutions $\Psi = \mu^l x$, $= \mu^m$ yield $l = 3/2$, $m = 1/4$, so that the width of the frictional sub-boundary layer is of the order $(10^{-2})^{1/4}$ or about one-fourth the width of the current. However, it can be argued that at this short

distance from the equator the vertical stresses come into play and the flow may no longer be regarded as two-dimensional. Indeed, if one accepts the values $\nu \sim 10\ \text{cm}^2\ \text{sec}^{-1}$ and $A \sim 10^6\ \text{cm}^2\ \text{sec}^{-1}$, one finds that near the equator the vertical eddy stress terms in the equations of motion dominate the lateral stress terms, though by no more than an order of magnitude.

We have derived inter alia the estimate $\delta = (U/\beta)^{1/2} \sim 200$ km for the order of magnitude of the half-width of the Cromwell Current. This estimate was obtained by equating orders of magnitude of the field and Coriolis accelerations in the equations of motion, and is therefore independent of the assumptions concerning the two-dimensionality of the motion. It is in fact consistent with our earlier estimate (eqs. 3.3) if one takes into account that the estimate, $u \sim (P^2/\beta)^{1/3} \sim 80$ km, is too small because of the untenable assumption that the inertial terms dominate the frictional terms. Thus δ may be written $(P^2/\beta)^{1/3} \div \beta)^{1/2}$ and if $(P^2/\beta)^{1/3}$ is replaced by U we obtain the correct expression.

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Vol 6 1959

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The general thermal circulation in equatorial regions

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Abstract—An approximation valid near the equator is developed for a model of the general thermal circulation. The existence of equatorial effects is related to the vanishing of the Coriolis parameter, and the equatorial approximation is related to the mid-latitude circulation. Qualitative features of the solutions are discussed.

1. INTRODUCTION

A recent study (ROBINSON and STOMMEL, 1959) has been made of a fluid system on the β-plane, thermally driven by imposed boundary temperatures. The surface temperature imposed varied with latitude only. The fluid considered, although inviscid, possessed a vertical coefficient of heat conductivity. The system was bounded on the east by a meridional plane, and was of infinite depth. This model was proposed as related to the general thermal circulation of the oceans.

In the previous study, solutions for the non-linearly coupled temperature and velocity fields were obtained, the validity of the solutions being limited to mid-latitude regions away from the eastern coast. In particular, the solutions became singular as they approached the equator. It is the purpose of the present work to remove this singularity by developing an approximation appropriate to the equatorial region. To do so, it is necessary to restate the assumptions which define the model in a more formal manner. This allows additional implications to be deduced. In addition, the inviscid but conducting condition is relaxed. It is replaced by the more physically plausible assumption of a unit eddy Prandtl number.

(2) THE MID-LATITUDE APPROXIMATION

Since the physical basis of an approximation scheme lies in order of magnitude considerations, non-dimensional variables and parameters are introduced. In the following, primed variables have dimensions and unprimed do not. x' is positive eastward, y' northward, and z' upward. p' is the actual pressure minus the hydrostatic pressure due to the mean density ρ_0. The mean temperature of the surface is T_0, and the asymptotic value of the temperature is zero. The vertical eddy conductivity, the vertical eddy viscosity, the thermal expansion coefficient, a typical horizontal dimension, the magnitude of the earth's rotation, and gravitational acceleration are denoted respectively by κ, ν, α, L, Ω, g.

Let

$$(x',\ y',\ z') = L\,(x,\ y,\ z)$$

$$(u',\ v',\ w') = \left[\alpha T_0\, g\,(2\Omega)^{-1}\right](u,\ v,\ w)$$

$$P' = \left[\rho_0\, L\alpha T_0\, g\right] P$$

$$T' = T_0\, T$$

Then, under the Boussinesq approximation, the equations of conservation of momentum, mass, and heat appropriate to this model are

$$-n\sigma\epsilon\, u_{zz} - fv + p_x + n(uu_x + vu_y + wu_z) = 0 \tag{1}$$

$$-n\sigma\epsilon\, v_{zz} + fu + p_y + n(uv_x + vv_y + wv_z) = 0 \tag{2}$$

$$-n\sigma\epsilon\, w_{zz} - T + p_z + n(uw_x + vw_y + ww_z) = 0 \tag{3}$$

$$u_x + v_y + w_z = 0 \tag{4}$$

$$-\epsilon T_{zz} + (uT_x + vT_y + wT_z) = 0 \tag{5}$$

where

$$\epsilon = K2\Omega(\alpha T_0 gL)^{-1}, \quad n = \alpha T_0 g(4\Omega^2 L)^{-1}, \quad \sigma = \nu\kappa^{-1}$$

and f is the Coriolis parameter divided by 2Ω, a smooth* function of y.

Although full non-dimensional variables have been introduced, the variables will not in general be order unity quantities, except for the horizontal coordinates and the temperature. Amplitudes and scales will depend on the three parameters ϵ, n, σ.

For $T_0 = 25\ C^\circ$, $L = 10^8\ cm$, $\Omega = \cdot 7 \times 10^{-4}\ \text{sec}^{-1}$, $g = 10^3\ \text{cm-sec}^{-2}$,

$\alpha = 2 \times 10^{-4}\ (^\circ C)^{-1}$, $\nu = \kappa = 1\ \text{cm}^2\text{-sec}^{-1}$,

$\epsilon = 3 \times 10^{-13}$, $n = 2$, $\sigma = 1$.

ϵ is extremely small while n and σ are of order unity. Therefore ϵ will have the major effect in setting all scales and amplitudes. To exploit the smallness of ϵ in order to derive approximate equations, it is necessary to use the techniques of singular perturbation theory. In ordering the equations by this method, the amplitudes and scales of all quantities are initially determined, that is, before the integration of the still mathematically complicated approximate equations.

Proceeding, we introduce explicitly the ϵ-dependence of all the variables. This is done in terms of as yet unknown exponents, viz. $\zeta = \epsilon^a z$, $u = \epsilon^b U$, $v = \epsilon^c V$, $w = \epsilon^d W$, $p = \epsilon^e P$. These new variables are substituted into equations (1) – (5), and it is assumed that all variables approach close to their asymptotic ($z = -\infty$) values within unit range of ζ, and that all dependent variables are smooth functions of x, y, ζ. Thus all the ϵ-dependence of each term of each equation is now exhibited explicitly in the coefficient of the term. The five exponents are determined by the following five assumptions:

(1) and (2) In equations (1) and (2), the pressure gradient and Coriolis terms are of the same order in ϵ.

(3) In equation (3), the pressure gradient and the buoyancy terms are of the same order in ϵ.

(4) In equation (4), the vertical divergence is the same order as the horizontal divergences.

(5) In equation (5), the vertical diffusion and the vertical advection are the same order in ϵ.

These five conditions yield $a = -1/3$, $b = c = e = 1/3$, $d = 2/3$.

*i.e., the scale of y is order of unity, and $f(y)$ itself and all its derivatives are upper bounded by unity.

Vol 6 1959

In terms of these exponents, the coefficient in ϵ of each term in all five equations can be computed. The equations can now be expanded by keeping in each only the lowest order ϵ-terms. The resulting approximate equations are

$$-fV + P_x = 0 \tag{6}$$

$$fV + P_y = 0 \tag{7}$$

$$-T + P_\zeta = 0 \tag{8}$$

$$U_x + V_y + W_\zeta = 0 \tag{9}$$

$$-T_{\zeta\zeta} + UT_x + VT_y + WT_\zeta = 0 \tag{10}$$

It should be noted that n and σ do not appear in these equations, so that to this approximation, all quantities of interest are independent of these parameters. In the momentum equations the amplitude ratios in ϵ, of the first neglected terms compared to the retained terms are:

(1) $O(\epsilon^{1/3})$, viscous and inertial terms

(2) $O(\epsilon^{1/3})$, viscous and inertial terms

(3) $O(\epsilon)$, viscous and inertial terms

Note also that all terms are important in the continuity and heat equations even to the lowest order of this approximation, and that the system is much more highly hydrostatic than geostrophic.

Equations physically identical to (6)–(10) were the assumed starting point of the previous study, and a full discussion of their solutions is given in that paper. The formal discussion of the assumptions leading to these equations given here serves two purposes. Firstly, the relative size of neglected effects can be numerically estimated i.e., a range of validity of the theory in ϵ is obtained. Secondly, in regions where the solutions of these approximate equations break down, the nature of the breakdown can be related to the assumptions violated. An appropriately modified approximation scheme can then be derived.

(3) THE EQUATORIAL APPROXIMATION

The behaviour of the solutions obtained from equations (6)–(10) is such as to invalidate their application near the equator. There the thermocline depth approaches zero, and all fields approach zero, except for W which diverges. In particular the asymptotic value of the vertical velocity, W_∞, is inversely proportional to the two-thirds power of the non-dimensional Coriolis parameter,

$$W_\infty \propto f^{-2/3} \tag{11}$$

Thus as $f \to o$, W is no longer a smooth function of y. Terms involving W in our original equations (1)–(5) may no longer be negligible compared to terms retained in equations (6)–(10).

This lack of smoothness is associated with amplitude changes, which must be considered in obtaining the approximation appropriate to the region about $f = 0$.*

*The mathematical technique which is being employed here may be found in an article by CARRIER (1954).

Furthermore, in addition to allowing for new ϵ-dependence of quantities whose amplitudes and scales were found in the mid-latitude regions, a new quantity of primary interest enters. This is the scale of y in which equatorial effects occur, i.e. the width of the singular region. As above. we introduce . .

$$\eta = \epsilon^A y, \quad \xi = \epsilon^B z, \quad \mu = \epsilon^C u, \quad \nu = \epsilon^D v, \quad \omega = \epsilon^E w, \quad p = \epsilon^F \varpi,$$

substitute in equations (1)–(5), and assume μ, ν, ω, ϖ, T to be smooth functions of x, η, ξ. The following set of assumptions is used to determine the six exponents and thereby to order the equations in the equatorial region.

(1) In equation (1), the vertical momentum advection and the pressure gradient terms are of the same order in ϵ.

(2) In equation (2) the pressure gradient and Coriolis terms are of the same order in ϵ.

(3) In equation (3), the pressure gradient and the buoyancy terms are of the same order.

(4) All terms in the continuity equation, (4), are of the same order in ϵ.

(5) In equation (5), the vertical diffusion and vertical advection are the same order.

(6) In the region where the mid-latitude and equatorial approximations overlap in validity, the same amplitude of W must be given by the two approximations.

Assumptions (1) and (5) are requirements consistent with the apparent equatorial divergence of the vertical velocity. The form of (1) is also consistent with the mathematical requirement that some term from the old approximation remain to balance the newly required term. This is necessary in order that the same solution to the original full equations may be obtained from the two approximations valid in different regions. p_x is chosen instead of fv because of the equatorial asymmetry of v. Assumption (2) differs from (1) because $\partial/\partial y$ becomes large as well as w, because of the equatorial symmetry of u and p, and because of the anticipation that a large linear term will dominate a large non-linear term. Assumption (4) is the only one containing D, the amplitude of v. Thus A, B, C, E, F are determined independently of (4), which is then the only assumption which yields five consistent approximate equations necessary for the five dependent fields.*

The joining condition, assumption (6), which relates amplitudes in the two regions by matching the ϵ-dependence of w at the as yet undefined 'edge' of each region, is mathematically realized as follows. Approaching the region of overlap from the equator, we recall that everywhere in the equatorial region $w = \epsilon^E \omega$. Since ω is smooth in η, $\omega_\infty = \omega(\eta \to \infty)$ is independent of ϵ; furthermore, ω_∞ is closely approached in order unity range of η. Approaching from mid-latitudes, we insist that when y becomes sufficiently small so that η is of order unity, the amplitude of w given by the mid-latitude approximation must also be ϵ^E. Taking the origin of y and η at the equator, let $f = \beta y = \beta \epsilon^{-A} \eta$, where β is 0 (1). Using equation (11), and recalling that in mid-latitudes $w = \epsilon^{2/3} W$, assumption (6) is found to require that

*This paragraph is, of course, a discussion and not a justification of our assumptions. The assumptions are a formal way of stating our physical model. The only logical justification lies in the internal consistency of the ordered equations implied by the assumptions. This consistency must be examined *a posteriori*. Any real justification lies in a comparison with nature.

Vol 6 1959

$\epsilon^{2/3}\,\epsilon^{2/3A} = \epsilon^{E}$, or $\frac{2}{3}(1 + A) = E$. Together with (1)–(5), the exponents are determined as

$$A = -\tfrac{1}{10},\quad B = -\tfrac{2}{5},\quad C = \tfrac{1}{5},\quad D = \tfrac{3}{10},\quad E = \tfrac{3}{5},\quad F = \tfrac{2}{5}.$$

In terms of these exponents, the equations expanded to lowest order take the approximate forms,

$$-\sigma n\mu_{\xi\xi} - \beta\eta v + \varpi_x + n(\mu\mu_x + v\mu_\eta + \omega\mu_\xi) = 0 \tag{12}$$

$$\beta\eta\mu + \varpi_\eta = 0 \tag{13}$$

$$-T + \varpi_\xi = 0 \tag{14}$$

$$\mu_x + v_\eta + \omega_\xi = 0 \tag{15}$$

$$-T_{\xi\xi} + (\mu T_x + vT_\eta + \omega T_\xi) = 0 \tag{16}$$

Thus the equations for the new scaled variables differ from the old approximate equations only in the first momentum equation. There, all non-linear terms as well as the frictional term are necessarily of the same order of magnitude as the geostropic terms. Compared to the retained terms the neglected terms in the second and third momentum equations are

(2) $O(\epsilon^{1/5})$, viscous and inertial terms.

(3) $O(\epsilon^{4/5})$, viscous and inertial terms.

Thus the system is less geostrophic and hydrostatic than it was at mid-latitudes.

Furthermore, for a system which is driven by wind stress as well as thermal forces, the appearance of the viscous term in (12) is important. It implies a much stronger coupling of wind-driven and thermal circulations than occurred at mid-latitudes. Thus in this region we can no longer relate our model to a real ocean by imposing a non-zero vertical velocity at the surface, i.e. by imposing the divergence of a wind-driven Ekman layer. The additional boundary condition required by the higher order z-term in (12) may be taken as a specification of the surface stress. Specification of zero surface stress now corresponds to a purely thermally-driven system; however, to make any realistic comparison with a real ocean, a non-zero stress must be applied.

Equations (12)–(16) are of course independent of ϵ, and serve to determine the form of μ, v, ω, T, ϖ, as well as the amplitude dependence of these fields on η and σ. However, as mentioned above, since (6)–(12) and their boundary conditions contain only order unity parameters and variables, they can yield only order unity amplitude factors. Thus, in obtaining the ϵ-dependence of the amplitudes and scales we have in fact obtained the numerical order of magnitude of the variables. Using the numerical values given in section (1), we obtain: the scale of equatorial effects as the order of 100 km, the magnitude of the east–west velocity in the equatorial region as the order of 100 cm sec^{-1}, and the equatorial thermocline as almost an order of magnitude shallower than at mid-latitudes. In addition, the equatorial amplitudes of the vertical and north–south velocities are found to be slightly greater than at mid-latitudes, $O(\epsilon^{.60})$ compared to $O(\epsilon^{.67})$, and $O(\epsilon^{.30})$ compared to $O(\epsilon^{.33})$, respectively.

(4) DISCUSSION

That the main features of the Cromwell current may indeed be related to the general thermal circulation is apparent from the above numerical results. The Cromwell current is observed to have a maximum amplitude of about 150 cm sec^{-1}, and

a width of about 200 km. The dense packing of isothermal surfaces near the equator obtained from our analysis is also compatible with observations on real oceans. However to make a meaningful comparison of our model with a real ocean, at least one additional qualitative result is needed. This is the sign of the strong equatorial current, or the latitudinal slope of the isothermal surfaces just north or south of the equator. This information is contained in equations (6)–(12) and in their boundary conditions. Since these equations are highly non-linear and involve many coupled variables, it is difficult to obtain even the gross features of their solutions.

However, dimensional equations physically equivalent to (6)–(12) form the assumed starting point for a study of the equatorial undercurrent by VERONIS (1960). In a paper appearing elsewhere in this issue, power series solutions of these equations in the latitudinal coordinate are developed about the equator. These solutions are for a longitudinally varying imposed surface temperature distribution. Associated with this boundary condition, a special type of longitudinal dependence is allowed, viz. u', p', T' are linear in x', and v' w' are independent of x'. The solutions obtained do contain an eastward flowing Cromwell current. Veronis' results do not strictly apply to our problem because of the different surface temperature condition imposed, but it is possible to reformulate simply our original problem to eliminate this difference. We consider therefore the general thermal circulation subject to the condition that $T(x, y, o) = T_0(y) + T_1 x$ where $T_0(y)$ is a prescribed function and T_1 a prescribed constant. Considering first the fields at mid-latitudes, solutions are sought with the special x-dependence stated above, rather than with the x-dependence given by our previous similarity transformation. In this case it can easily be shown that the asymptotic value of the vertical velocity remains inversely proportional to the two-thirds power of the Coriolis parameter, equation (11) is maintained. Since this is the feature of the mid-latitude solution used in deriving the equatorial approximation, all the results of section (2) are unaltered, including numerical results. Thus a qualitative resemblance, both in order of magnitudes and in signs, is seen to exist between the results of our model and oceanic observations.

Emphasis has been placed on the gross features of the solution to these equations not only because of the mathematical difficulty of obtaining exact solutions, but also because a comparison is made with average observations on a real ocean. The crude model proposed, containing constant eddy parameters, can in no sense be expected to bear a precise quantitative relation to the fundamentally turbulent geophysical situation. It is, however, meaningful to make a qualitative or semi-quantitative comparison. What has been studied is the effect of the driving forces and rotational constraints on the mean fields. The turbulent transfer processes are essential and must be included. In terms of eddy parameters, they are included in a crude and hypothetical way. The approach is a meaningful one only because there is a large number of terms in our equations, other than the eddy parameter terms, which govern the gross features of the solutions.

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Vol 6 1959

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Vol.
6
959-60

An approximate theoretical analysis of the equatorial undercurrent

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INTRODUCTION AND SUMMARY

THE EQUATORIAL Undercurrent in the Pacific Ocean is a narrow (2° wide) subsurface current flowing to the east. The prevailing winds in this equatorial region are toward the west. CROMWELL, *et. al.*, (1954) made the first oceanographic observations of the current in 1952. KNAUSS (1960) later undertook a more extensive observational programme and confirmed the original observation. The current has a maximum eastward velocity of 150 cm/sec at a depth of about 100 m. At the surface and at depths greater than 300 m the flow is relatively weak and toward the west.

A simple physical argument can be put forth to explain the existence of the undercurrent. A wind blowing westward over a non-rotating, bounded basin will pile up water on the western ride and thus give rise to a pressure gradient which drives the subsurface water back toward the east (Fig. 1). When the water is vertically

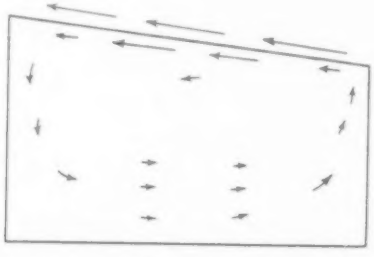

FIG. 1. A sketch of the overturning cellular motion in a confined non-rotating basin when the surface is subjected to a tangential shear stress acting to the left. The pile-up of water toward the left forces a slow return circulation in the deeper parts of the basin.

stratified, the depth to which the current extends may be limited by the adjustment of the stratified layers. At the equator where the effect of the earth's rotation is essentially negligible the above picture obtains. In regions away from the equator the Coriolis force becomes appreciable and, indeed, dominates the system.

The present paper is a study of the dynamics of the system in the immediate vicinity of the equator. The investigation is restricted to an analysis of the vertical structure of the current. No attempt has been made to determine the parameters which control the width of the current. ROBINSON (1960) has investigated this phase of the problem and has, by dimensional arguments, deduced a width of approximately 100 km. The system of equations which he deduces as the equations which control the motion, coincide with those treated in this paper and which are arrived at by a

physical argument. He also shows that this system of equations joins with those of the thermocline model of ROBINSON and STOMMEL (1959) at about 1° latitude.

The non-linear equations are treated approximately by replacing parts of the non-linear convective terms by average values.* For example, W, an average value of the vertical velocity, is substituted for w in the terms $w\frac{\partial T}{\partial z}$ and $w\frac{\partial u}{\partial z}$. The quantity W is determined by averaging the final solution for w.

In addition the various quantities are expanded in a power series in y, the north–south coordinate. Solutions are derived for the coefficients of y^0, y^1 and y^2 only, since the essential behaviour of the fluid system near the equator is then immediately available. A final simplification is the choice of a very specific dependence of the variables on the east-west co-ordinate, x. Essentially, the latter implies a treatment of the system away from coastal or other barriers at points along the equator. A boundary-layer solution would have to be added to satisfy boundary conditions at the east-west boundaries.

The model treated requires a knowledge of the temperature (or the heat flux) at the surface of the ocean. The method of solution allows one to fix only that part of the surface temperature which is independent of y. The part of the surface temperature which varies with y is then deduced. In one sense this inability to fix the surface temperature completely is desirable since the final results can then represent a test of the model.

The first conclusion from this analysis is that the depth of penetration of the current system is proportional to $\sqrt{-\frac{\tau L}{g\alpha\bar{T}}}$, where τ is the east – west wind-stress (positive toward the east), L is the east-west scale of the system, g is the gravitational acceleration, α is the coefficient of thermal expansion and $\bar{T}$ is the difference in surface temperatures between the western and eastern edges of the system. It is important to note that the above expression is independent of any internal parameters such as eddy coefficients etc. Using $\tau = -0{\cdot}2$ cm²/sec*, $L = 10^9$ cm, $g = 10^3$ cm/sec², $\alpha = 2{\cdot}5 \times 15^{-4}$/°C and $\bar{T} = 5$ °C†, one finds that the scale depth is 125 m. This is also the depth of the core of the undercurrent and it agrees reasonably well with Knauss' measurement of 100 m.

Fig. 2 shows a vertical profile of the east-west current at the equator (u is independent of longitude in this analysis). The *shape* of the profile is given by the present theory but the *surface value* depends strongly on the depth over which the vertical averages for W etc. are made. In Fig. 2 the surface value has been chosen as — 15 cm/sec. The remaining values are then completely determined. The agreement with Knauss' measurement is quite good. The maximum velocity of the undercurrent is 170 cm/sec toward the east as compared to the observed value of 150 cm/sec.

In Fig. 3 the temperature at the eastern and western extremities along the equator is plotted as a function of the vertical co-ordinate z. The absence of a sharp thermocline near the equator appears in the observations also. However, the water near

*One cannot expect precise quantitative results with this model unless the non-linear terms are actually unimportant. However, in that case the equations could be linearized from the outset.

†The value $\tau = -0{\cdot}2$ cm²/sec was suggested by KNAUSS for the average value of the east-west wind stress component. The value $\bar{T} = 5$ °C was taken from Муромцев, А. М., 1958, Основные Черты Гидрологии Тихого Океана, Приложениед, Атлас Вертикальных Разрезов и Карт Температуры, Солености, Плотности и Содержания Кислорода, Ленинград.

the surface (upper 100 m) is more nearly water of constant temperature than one deduces from this analysis.

An east-west profile of temperature vs. depth is shown in Fig. 4. The qualitative agreement with observation is again reasonable considering the restrictions imposed by the specific east-west dependence chosen here.

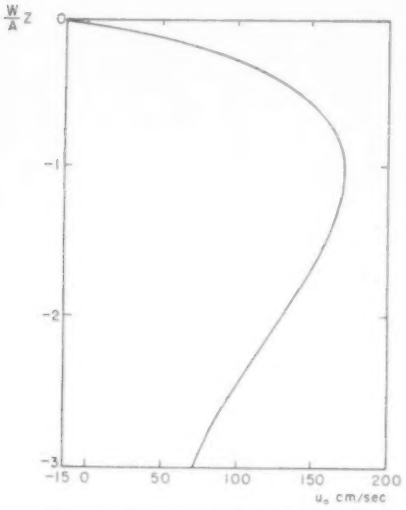

FIG. 2. The vertical profile for u_0.

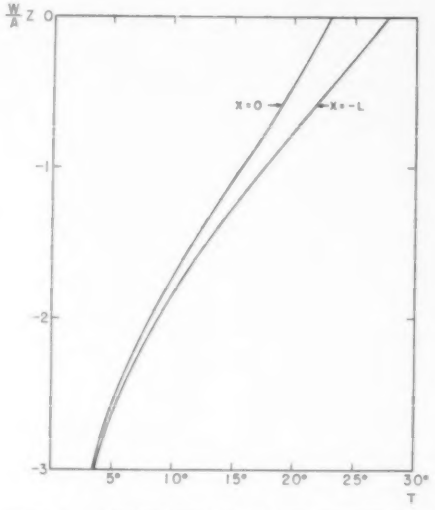

FIG. 3. Vertical profiles of the temperature T_0 at $x = 0$ and at $x = -L$.

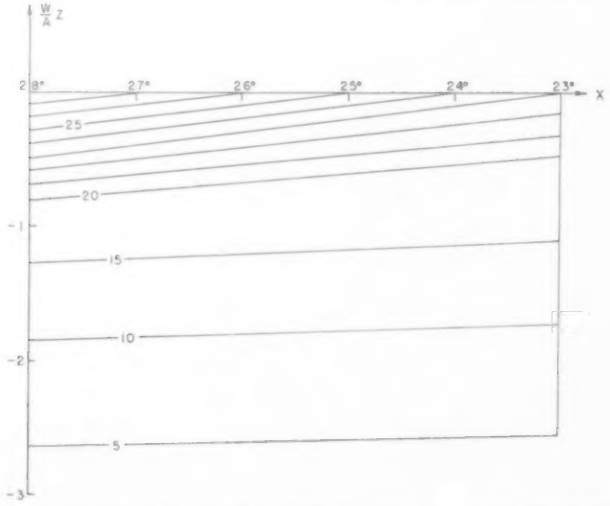

FIG. 4. A vertical cross-section along the equator of the theoretically determined temperature distribution.

One cannot extend the analysis to include much more quantitative information than that discussed up to this point. Qualitatively, one can see that the temperature above the core increases with latitude and the temperature below the core decreases.

This gives rise to the relative vertical spreading of the isotherms at the equator. For further quantitative results one would have to include the east-west convective terms in the second order equations but, as discussed in the next section, the iterative approach hinges on an averaging process which cannot lead to quantitative results.

It should be emphasized that the model described here is oriented toward solving a specific problem. One can see this most readily if one tries the same approach for the purely thermal model of ROBINSON and STOMMEL (1959) near the equator. Since the prescribed surface temperature is intimately connected with the presence of a wind-stress at the surface, one cannot impose the same simple east-west dependance of the variables for the analysis of the purely thermal equatorial problem.

There are other approaches to this problem which are oriented toward explaining features other than the vertical structure. FOFONOFF and MONTGOMERY (1954) have shown that the existence of the undercurrent is consistent with conservation of potential vorticity. STOMMEL (1960) and CARRIER (1960) each look at the undercurrent in the light of the Ekman layer, near the equator. LONG (1960) treats the undercurrent as one example of eastward moving jets in the atmosphere-ocean system. ROBINSON'S (1960) approach has already been mentioned. Finally, there is the ever-present problem of the stability of the undercurrent. What are the factors which limit the current to a 2° band centred on the equator? C. C. LIN, during a seminar in which the undercurrent was discussed at M.I.T., suggested the interesting possibility that an eastward moving jet may be stabilized by the Coriolis effect. He notes that if the jet is pushed toward the north or south, the Coriolis force acts on the current to return it to its equatorial position. For a westward moving jet the same force causes it to become unstable.

ANALYSIS

The system of equations which is taken to be descriptive of the Equatorial Undercurrent is

$$u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + A\frac{\partial^2 u}{\partial z^2} \tag{1}$$

$$fu = -\frac{1}{\rho}\frac{\partial p}{\partial y} \tag{2}$$

$$\frac{1}{\rho}\frac{\partial p}{\partial z} = g\alpha T \tag{3}$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \tag{4}$$

$$u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + w\frac{\partial T}{\partial z} = A\frac{\partial^2 T}{\partial z^2} \tag{5}$$

where x, y, z are co-ordinates which are positive eastward, northward and upward respectively; α is the coefficient of thermal expansion. The constants A in (1) and (5) is an eddy coefficient. The assumption is that heat and momentum mixing is generated by the same type of small scale motion and that the effects of this mixing can be parameterized by a constant coefficient. The remaining variables are the usual ones.

This system of equations is based on the Boussinesq approximation where the variation of density $\rho = \rho_0(1 - \alpha T)$ is taken into account only in the gravitational term. Terms have been neglected not only on the basis of a formal expansion but simply on the basis of physical arguments. For example, the non-linear terms have been kept in equation (1) where at the equator the familiar geostrophic approximation fails completely. However, in equation (2) the neglected terms vanish at the equator (as do the terms retained) and the geostrophic approximation is therefore kept as a reasonable first approximation. Horizontal mixing has been neglected throughout since attention is confined primarily to the vertical structure. An *a posteriori* check of the above equations shows that the neglected terms are never larger than 10 per cent of those kept.

The problem as it stands now is to solve the above system of equations together with appropriate boundary conditions. There are obvious difficulties in trying to solve this general problem and certain approximations will therefore be made.

First the vertical velocity, w, is replaced by an average value W = constant* in the convective terms. Equations (1) and (5) then become

$$u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + W\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + A\frac{\partial^2 u}{\partial z^2} \tag{6}$$

$$u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + W\frac{\partial T}{\partial z} = A\frac{\partial^2 T}{\partial z^2} \tag{7}$$

For the y-- dependence we assume symmetry about the equator for u, w, p and T and anti-symmetry for v. We can then write

$$(u, w, p, T) = \sum_0^\infty (u_{2n}, w_{2n}, p_{2n}, T_{2n})\, y^{2n} \tag{8}$$

$$v = \sum_0^\infty v_{2n+1}\, y^{2n+1} \tag{9}$$

The Coriolis parameter f will be written

$$f = \beta y \tag{9.a}$$

If (8), (9) and (9.a) be substituted into (2), (3), (4), (6), (7) and if the coefficients of the various powers of y be equated, the following system of equations obtains:

$$u_0\frac{\partial T_0}{\partial x} + W\frac{\partial T_0}{\partial z} = A\frac{\partial^2 T_0}{\partial z^2} \tag{10}$$

$$\frac{1}{\rho}\frac{\partial p_0}{\partial z} = g\alpha T_0 \tag{11}$$

$$u_0\frac{\partial u_0}{\partial x} + W\frac{\partial u_0}{\partial z} = -\frac{1}{\rho}\frac{\partial p_0}{\partial x} + A\frac{\partial^2 u_0}{\partial z^2} \tag{12}$$

$$\frac{\partial u_0}{\partial x} + v_1 + \frac{\partial w_0}{\partial z} = 0 \tag{13}$$

*This value of W must, of course, be deduced from the final solution and, therefore, it is not an 'available' parameter.

Vol 6 1959

$$\beta u_0 = -\frac{2}{\rho} p_2 \tag{14}$$

$$u_2 \frac{\partial T_0}{\partial x} + u_0 \frac{\partial T_2}{\partial x} + 2v_1 T_2 + W \frac{\partial T_2}{\partial z} = A \frac{\partial^2 T_2}{\partial z^2} \tag{15}$$

$$\frac{1}{\rho} \frac{\partial p_2}{\partial z} = g\alpha T_2 \tag{16}$$

$$u_0 \frac{\partial u_2}{\partial x} + u_2 \frac{\partial u_0}{\partial x} + (2u_2 - \beta) v_1 + W \frac{\partial u_2}{\partial z} = -\frac{1}{\rho} \frac{\partial p_2}{\partial x} + A \frac{\partial^2 u_2}{\partial z^2} \tag{17}$$

$$\frac{\partial u_2}{\partial x} + 3v_3 + \frac{\partial w_2}{\partial z} = 0 \tag{18}$$

$$\beta u_2 = -\frac{4}{\rho} p_w \tag{19}$$

etc.

Now the surface temperature at the equator is approximately a linear function of x. This suggests a substitution of the form

$$T = T_{00} + T_{01}\, x, \qquad p_0 = p_{00} + p_{01}\, x \tag{20}$$

where the form of p_0 comes from the hydrostatic relation. The remaining variables can be chosen consistently to be independent of x (since the wind-stress is also taken independent of x). One can look upon the present restricted form for the variables as appropriate away from the coastal boundaries. To satisfy such conditions as zero flow through the x– boundaries, it is necessary to add a boundary-layer solution to the present system.

We now propose to solve the equations to order y^2 with the specific x– dependence described above and with the boundary conditions

$$z = 0: \qquad T_0 = \bar{T}\left(c - \frac{x}{L}\right), \qquad A \frac{\partial u_0}{\partial z} = \tau = \text{constant}$$

$$z = \infty: \qquad T_i = u_i = v_i = p_i = 0$$

The condition on T_0 at the surface simply prescribes the surface temperatures as a linear function of x. In the Equatorial Pacific $\bar{T} \approx 5\,^{\circ}\text{C}$ and $c\,\bar{T}$, the temperature at the eastern edge ($x = 0$) of the ocean is abour 23°. Therefore $C = 4{\cdot}6$.

Substituting (20) into (10), one has

$$u_0 T_{01} + W \frac{\partial T_{00}}{\partial z} = A \frac{\partial^2 T_{00}}{\partial z^2} \tag{21}$$

$$W \frac{\partial T_{01}}{\partial z} = A \frac{\partial^2 T_{01}}{\partial z^2} \tag{22}$$

Equation (12) becomes (since $\frac{\partial u_0}{\partial x} = 0$)

$$W \frac{\partial u_0}{\partial z} = -\frac{1}{\rho} \frac{\partial p_0}{\partial x} + A \frac{\partial^2 u_0}{\partial z^2} \tag{23}$$

Integrating (22) and applying the condition that $T_{01} \to 0$ as $z \to -\infty$ gives

$$T_{01} = -\frac{\bar{T}}{L} e^{\frac{W}{A}z} \tag{24}$$

where $-L \leqslant x \leqslant 0$ and T is the total amplitude of the east west temperature variation ($\approx 5°$ at the equator in the Pacific).

From (11)

$$\frac{p_{01}}{\rho} = -\frac{g\alpha\bar{T}A}{LW} e^{\frac{W}{A}z} \tag{25}$$

where we have taken $p_{01} \to 0$ as $z \to -\infty$.

Substitution (25) into (23), we have as a solution for u_0,

$$u_0 = \frac{\tau}{w}\left[1 + \frac{g\alpha T}{L\tau}\frac{A^2}{W^2}\left(1 - \frac{W}{A}z\right)\right] e^{\frac{W}{A}z} \tag{26}$$

where the conditions

$$A\left.\frac{\partial u_0}{\partial z}\right|_{z=0} = \tau = \text{constant}, \qquad \left.u_0\right|_{z=-\infty} = 0$$

have been used. In the Pacific $\tau \approx -0{\cdot}2\ \text{cm}^2/\text{sec}$ near the equator. The quantities u_0 and T_{01} can be substituted into (21) to yield T_{00} which is

$$T_{00} = 4\,T\,e^{\frac{W}{A}z} + \frac{\tau TA}{2LW^2}\left\{\left(1 + \frac{5}{2}\frac{g\alpha T}{\tau L}\frac{A^2}{W^2}\right)\left(e^{\frac{W}{A}z} - e^{2\frac{W}{A}z}\right) + \frac{g\alpha T}{\tau L}\frac{A}{W}\,z\,e^{2\frac{W}{A}z}\right\} \tag{27}$$

In (27) the condition $T_0 = 4{\cdot}6\,T$ at $z = o$, $x = o$ has been satisfied. This condition merely states that the surface temperature at the eastern edge of the basin is 23 °C. Equations (24) and (27) which together give the temperature at the equator satisfy the condition that the surface temperature at the equator varies between 23° at the eastern edge and 28° at the western edge.

From (14) and (16)

$$g\alpha T_2 = -\frac{\beta}{2}\frac{\partial u_0}{\partial z}$$

or

$$T_2 = \frac{\beta\bar{T}}{2LW}\left(-\frac{\tau LW}{g\alpha TA} + z\right) e^{\frac{W}{A}z} \tag{28}$$

The second order temperature, T_2, is automatically prescribed everywhere and one cannot satisfy any boundary conditions for T_2. We shall propose therefore to investigate the surface values of T_2 in the final solution and if the values deviate very much from the observed boundary values, the present method of solution must be discarded. We shall see, however, that fairly reasonable results are deduced from (28).

To proceed one must solve equations (15) and (17) for u_2 and v_1 with $\frac{\partial p_2}{\partial x} = 0$, $\frac{\partial T_2}{\partial x} = 0$, $\frac{\partial u_2}{\partial x} = 0$, $\frac{\partial u_0}{\partial x} = 0$. The second of these is non-linear and no general solution to the system has been found. Approximate solutions can be found in the following way:

Vol 6 1959

Replace $2T_2$ by an average value $\theta = \text{const.}$ and $2U_2$ by an average value $s = \text{const.}$ Furthermore $u_2 \frac{\partial T_0}{\partial x}$ will be taken into account by an iteration procedure, i.e., first one solves

$$\vartheta v_1 = A \frac{\partial^2 T_2}{\partial x^2} - W \frac{\partial T_2}{\partial x} \tag{29}$$

for v_1, then

$$(S - \beta) v_1 = A \frac{\partial^2 u_2}{\partial z^2} - W \frac{\partial u_2}{\partial z} \tag{30}$$

for u_2. One can then substitute the u_2 solution into the term $u_2 \frac{\partial T_0}{\partial x}$ of (15), solve for a corrected v_1 which can replace the old v_1 in (17) to correct u_2 etc.

The first step in this procedure gives

$$v_1 = \frac{\beta T}{2L\vartheta} e^{\frac{W}{A}z} \tag{31}$$

and

$$u_2 = -\frac{(s - \beta)\beta T}{2L\vartheta W}\left(\frac{A}{W} - z\right) e^{\frac{W}{A}z} \tag{32}$$

where $\frac{\partial u_2}{\partial z} = 0$ at $z = 0$.

Rather than proceed with the iteration procedure we can return to (13) to solve for w_0 using (31) for the value of v_1. This yields

$$w_0 = \frac{\beta T A}{2L\vartheta W}\left(1 - e^{\frac{W}{A}z}\right) \tag{33}$$

where $w_0 = 0$ at $z = 0$.

If one now returns to the definition of W one can deduce W from (33). Integrating (33) from a depth $z = -\frac{cA}{W}$ (where c is some constant) to $z = 0$, one has

$$W^2 = \frac{\beta T}{2L\vartheta C}\left[c - (1 - e^{-c})\right] \tag{34}$$

Similarly integrating T_2 in (28) from $-\frac{cA}{W}$ to 0,

$$\vartheta = -\frac{\beta\tau}{g\alpha Ac}(1 - e^{-c}) - \frac{\beta T A}{LcW^2}\left[1 - (1 + c)\, e^{-c}\right] \tag{35}$$

Solving (34) and (35) jointly yields

$$\frac{W^2}{A^2} = -\frac{g\alpha T}{2L\tau}\frac{\left[c + 1 - (1 + \alpha c)\, e^{-c}\right]}{1 - e^{-c}} \tag{36}$$

$$\vartheta = -\frac{\beta\tau}{g\alpha A}\frac{(1 - e^{-c})(c - 1 + e^{-c})}{c\left[c + 1 - (1 + 2c)\, e^{-c}\right]} \tag{37}$$

From (36) and (37) one can immediately see the qualitative dependence of θ and W/A (the reciprocal of the scale depth) on the given parameters of the problem. It is interesting to note first that A/W is independent of all internal parameters of the

problem. Specifically, the scale depth does not depend on the internal mixing parameter, A. The remaining parameters g, α, T, L, τ are all well defined (τ may be difficult to deduce quantitatively in practice but the only pertinent point is that it has a single value).

Before proceeding further I should like to mention a simple, qualitative argument proposed by L. N. Howard by which one can deduce the same parametric dependence for the scale depth. Consider a bounded non-rotating basin where the average temperature above a scale depth, h, varies from $T + \Delta T$ at the western edge to T at the eastern edge. If the horizontal pressure gradient vanishes at the depth, h, the level $h + \Delta h$, of the surface at the warmer end will be higher than that at the cooler end. The pressure at the depth $h + \Delta h$ in the warmer region is $(h + \Delta h)\ (\rho_0 - \Delta\rho)g$ where $\rho_0 - \Delta\rho$ is the average density of the water there. In the cooler region the corresponding pressure is $\rho_0 gh$. Since the horizontal pressure gradient vanishes $\rho_0 gh = (h + \Delta h)\ (\rho_0 - \Delta\rho)\, g$. Hence, to first order in $\Delta\rho$ and Δh, $\rho_0 \Delta h = h\Delta\rho$. At depths less than h the change in pressure along a horizontal surface is of order $\Delta p = \rho_0 g \Delta h = gh\Delta\rho = gh\rho_0 \alpha \Delta T$ where we have used the relation $\Delta\rho = \rho_0 \alpha \Delta T$. Thus

$$\frac{\partial p}{\partial x} = - gh\rho_0\, \alpha\, \frac{\Delta T}{L}$$

where $\Delta T/L$ approximates the horizontal temperature gradient $\partial T/\partial x$. If the horizontal pressure gradient is approximately balanced by the vertical mixing, we have

$$\frac{\partial p}{\partial x} = - gh\rho_0\, \alpha\, \frac{\Delta T}{L} \sim \rho_0\, A\, \frac{\partial^2 u}{\partial z^2}$$

Integrating from $z = h$ to $z = 0$, we get

$$- gh\alpha\, \frac{\Delta T}{L}\, h \sim A\, \frac{\partial u}{\partial z}\bigg|_{z=0} = \tau$$

where we assume $\dfrac{\partial u}{\partial z} = 0$ at $z = - h$. Thus

$$h^2 \sim \frac{- \tau}{g\alpha \Delta T/L} = \frac{- \tau L}{g\alpha \Delta T}$$

Returning to (36) we chose c so that the integration ranges over the depth of interest, viz., about 2 scale depths. The qualitative results are unaffected by the explicit value of c but the quantitative results are. Quite arbitrarily therefore we choose $c = 1{\cdot}5$ which gives the surface value $u_0 = - 16$ cm/sec. Other choices for c yield quite different surface values for u_0.

However, the scale depth A/W which has the value 125 m for $c = 1{\cdot}5$ changes only by 20 per cent when c is halved or doubled. The $u_0 vs. z$ profile is also qualitatively unchanged by changes in c.

The results presented thus far are discussed in the introduction and summary. We shall now discuss the possibility of refining these results by taking into account higher order terms and by extending the iteration procedure discussed in the paragraphs following equation (28).

In the first place it should be noted that T_2 as given by (28) goes through zero at the depth $z_0 = \dfrac{\tau LW}{g\alpha TA}$. Hence, the approximation introduced by taking the average

Vol 6 1959

$\theta = \text{const.}$ for $2T_2$ is not generally a good one. (Since the positive part of T_2 is of greater amplitude than the negative part, the approximation may be acceptable for the region $z > z_0$). It can be seen that by this procedure v_1 must always be positive, a fact which is clearly at odds with both observation and the physics of the model. Furthermore, since u_2 is determined from v_1, it is also of one sign. Therefore, the procedure leads to inconsistent information regarding the higher order velocities.

One cannot alter the qualitative results by the iteration procedure proposed to provide higher order corrections to v_1 and u_2. The only appropriate procedure is to solve the non-linear system consisting of equations (15) and (17) for v_1 and u_2. This I have not been able to do. However, it should be noted that the analysis leading to the expressions for p_0, T_0, u_0, T_2, p_2 is independent of the subsequent procedure. The only necessary information which requires a knowledge of the quantities v_1 and v_2 is the parametric dependence of A/W, an evaluation which does not depend on the *form* of u_2 and v_1. Furthermore, since one can derive the same parametric dependence by means of Howard's argument there is some justification for keeping the limited results obtainable from this model.

The calculation for the temperature $T_0 + y^2T_2$ at the surface shows that the temperature increases by 0·4° at a distance of 50 km north or south of the equator. (This distance represents the limit of validity of the second-order solution. Higher order terms must be taken into account for larger values of y). This value of the temperature increase is approximately twice the observed value.

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Evidence for an equatorial undercurrent in the Atlantic Ocean

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(1) INTRODUCTION

WITH the discovery of the Equatorial Undercurrent in the Pacific Ocean (CROMWELL, MONTGOMERY, STROUP, 1954), another most interesting current became known as part of the complicated structure of the water movements in the equatorial region. The existence of the Cromwell Current in the Pacific Ocean has been established as a permanent feature of the oceanic circulation, and subsequent observations have contributed to a more detailed knowledge of this relatively swift current in the close vicinity of the equator.

Not much attention, however, has been paid to the possible existence of a similar phenomenon in the Atlantic Ocean. The following note may help to provide some evidence that the Atlantic also has such a current. This evidence is based partly on some direct older observations and partly on the results obtained in 1946 from an analysis of oceanographic stations between 10 °N and 10 °S. The latter was never published. Since that time more oceanographic stations have been added in the equatorial region of the Atlantic and therefore, it would be worthwhile to include the more modern stations and to check the results obtained in 1946.

(2) EARLY DESCRIPTIVE LITERATURE

O. KRÜMMEL (1911) paid considerable attention to the details of the Equatorial Current System in the Atlantic following the analysis by Capt. HOFFMANN of ' The nine ten-degree fields ' in the Atlantic equatorial region. According to HOFFMANN (1911, p. 550) the west-flowing South Equatorial Current, which normally extends over the equator into the Northern Hemisphere, has two maxima. One is found just north of the equator between 0° and 2 °N, the other south of the equator between about 2 °S and 6 °S. Hence, the South Equatorial Current has two maxima ' streaks,' one in about 1 °N, the other in about 4 °S. At the equator, or a little south of it, the surface currents are at a minimum.

At another place, (KRÜMMEL, 1911, p. 562) in connection with the currents directly at the equator, reports in a footnote on observations of extraordinary direction. This note reads (in translation) :

> ' If, exceptionally, in March the doldrums from the north move to the equator or even into southern latitudes, the Guinea-Current follows immediately, and eastward currents dominate at the equator*. This happened among other examples in March 1893 and 1894. In 1894, the

*Probably, what was observed is not the Guinea-Current (eastern part of the Equatorial Counter Current) ' which follows immediately,' but the Undercurrent at the equator which breaks through to the surface when the wind stress vanishes in the equatorial region.

German bark *Kepler* observed on the short distance between 1·5 °S, 27·0 °W and 2·5 °N, 24·3 °W, from March 2nd through March 13th, daily eastward currents with an average of 23 nautical miles per day.'

The maximum speed was observed on March 4th with 45 nautical miles per day going to the east (C. PULS, 1895). It is of interest to note that this ship was in the central part of the Atlantic Ocean or even closer to the South American coast when it experienced this strong and continuous eastward drift.

Also observations of J. Y. BUCHANAN on board the *Buccaneer* indicate certain 'anomalies' just at the equator. At approximately 10 °W, weak westward surface currents were observed, whereas at a depth of 55 metres the water moved toward the southeast with a velocity of more than 1 knot (O. KRÜMMEL, 1911, p. 552). Although these old observations may not be as reliable as modern measurements, they can not be disregarded. They seem to indicate that in the Atlantic Ocean an eastward flow also exists under the equator that may come to the surface when the wind stress exerted by the southeast trades at the equator vanishes. Normally, the pure wind drift towards the west caused by the trade winds superimposes and masks the eastward flow at the surface.

Results obtained by A. SCHUMACHER (1940) in an analysis of his Atlantic Surface Current Charts also seem to indicate the existence of an eastward current in the western Atlantic close to the equator. SCHUMACHER calls attention to the 'abnormal' set of ships in the region between 30° and 35 °W just south of the equator, where eastward currents were observed from early December to the beginning of May. In the western Atlantic this is the time when the thermal equator moves south to the vicinity of the geographical equator, or even south of it. SCHUMACHER calls these currents 'backward sets,' and refers to E. RÖMER (1937). It seems that these observations (SCHUMACHER, 1940, p. 120) give some 'climatological evidence' of the Equatorial Undercurrent because, climatologically, this region near the coast of South America is the only one in the Atlantic where the thermal equator and the doldrums reach the geographical equator most of the time during northern winter or early spring.

(3) ADDITIONAL SUPPORTING EVIDENCE

Based on oceanographic observations that were available until 1946* in the Atlantic between 10 °N and 10 °S, the anomaly of dynamic height of the sea surface and other isobaric surfaces above the 600 dbar surface was computed. Assuming the 600 dbar surface to be level, the values give the topography of the various pressure surfaces. R. B. MONTGOMERY and E. PALMÉN (1940) used the 1000 dbar surface as a reference surface for their study of the Equatorial Counter Current; however, the results are not significantly changed with either choice. Preference was given to the 600 dbar level upon consideration of the results obtained by A. DEFANT (1941).

In Fig. 1, the topography of the sea surface in the equatorial region of the Atlantic Ocean is shown. It deviates slightly from A. DEFANT's (1941) results. The high pressure ridge belonging to the Equatorial Counter Current in 5 °N to 10 °N, approximately, as well as the high pressure ridge associated with the Guayana Current,

*This data was used in a Colloquium at the University of Hamburg in 1946 in connection with a talk on the Equatorial Counter Current and its dynamical structure. Part of it was published, G. NEUMANN, 1947).

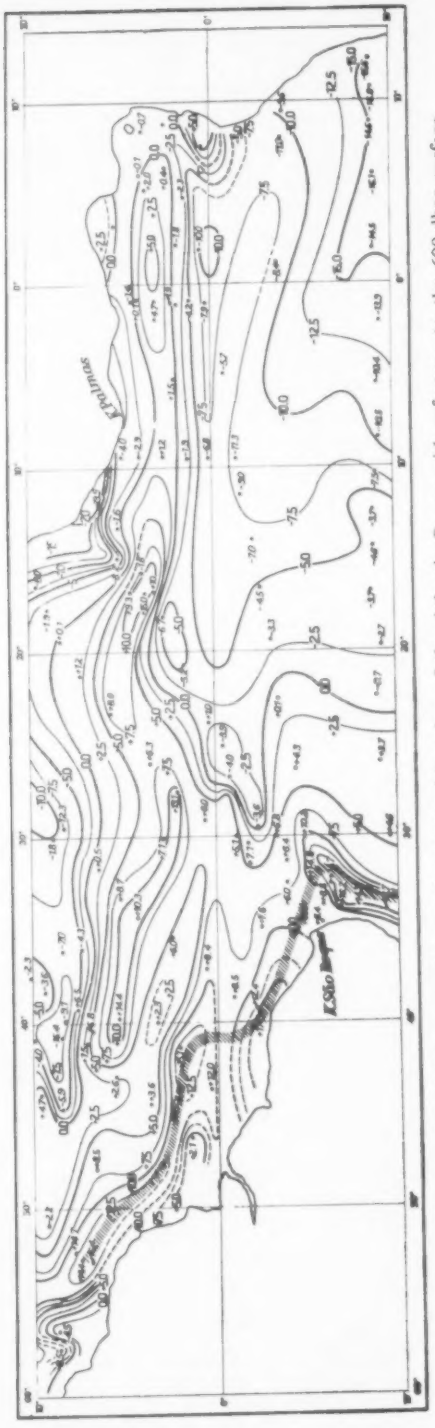

FIG. 1. Topography of the sea surface in the equatorial region of the Atlantic Ocean with reference to the 600 dbar surface.

Vol
6
1959

are clearly indicated. Along the equator there is a trough in which the sea surface slopes by about 22 cm from the western high pressure ridge down to a minimum around 2 °E. A few stations in the Gulf of Guinea seem to indicate a slope reversal at 1 °E or 2 °E, and a rise of the sea surface towards the African coast. The scarcity of complete oceanographic stations in this region, however, leaves some doubts about this rise to the east, although the temperature distribution at the 100-metre level seems to support this idea. Since more temperature observations close to the African coast were available than complete oceanographic stations (including salinity observations), Fig. 2 is shown presenting the distribution of temperature at a depth of 100 m. There is a strong indication that close to the coast of Africa, in the Gulf of Guinea, the temperature at a depth of 100 m rises again toward the east. This, in the absence of corresponding salinity observations, may at least serve to support the hypothesis that the sea level rises again close to the African coast.

Another characteristic feature of the temperature distribution in Atlantic equatorial regions may be mentioned in connection with figure 2, i.e., the temperature minimum that is observed at about 1 °S–2 °S. This region of relatively cold water just south of the equator was also a matter of KRÜMMEL's (1911, p. 554) special interest†. This temperature minimum just south of the equator is the result of the divergence in the wind driven currents as produced by the southeast trade winds. With winds blowing over the equator from the Southern Hemisphere a zone of maximum surface divergence develops just south of the equator. With winds blowing over the equator from the Northern Hemisphere, the zone of maximum divergence would develop just north of the equator (G. NEUMANN, 1947).

(4) ESTIMATE OF THE VELOCITY AT THE EQUATOR

For the development of the Equatorial Undercurrent, the zonal slope of the sea surface at the equator seems to be most important. This slope from west to east is strongest in the western part (Fig. 1), although some of the details shown on the chart may not be real, for example the extremely strong gradient at about 29 °W. The total of 22 dyn cm difference in the height of the sea surface over a level surface between west and east across the Atlantic agrees well with the 24 dyn cm difference obtained by MONTGOMERY and PALMÉN (1940).

If the Cromwell Current is associated with the piling up of the near surface waters on the western side of the Pacific Ocean by the Trade Winds, it should also exist in the Atlantic Ocean, since a similar strong slope of the sea surface is indicated from west to east as in the Pacific Ocean. In the Atlantic, one may expect to find an Equatorial Undercurrent which is strongest west of 20 °W longitude and weaker toward the east. It is even possible that close to the African coast a short undercurrent exists that flows to the west.

Based on the topographic chart for the sea surface in Fig. 1, a tentative gradient current field is drawn and presented in Fig. 3. It shows the Equatorial Counter Current between about 5 °N and 10 °N, and separated from it by a strong westward current is the eastward flow just at the the equator. While the Equatorial Counter Current, for its explanation, does not necessarily require a piling up of water against the western shores, but may exist even in an unbounded ocean provided that the trade winds are asymmetrically developed with respect to the equator (G. NEUMANN,

†For the Pacific Ocean, see KRÜMMEL, (1911) Vol. 1, page 410, Figs. 55–56.

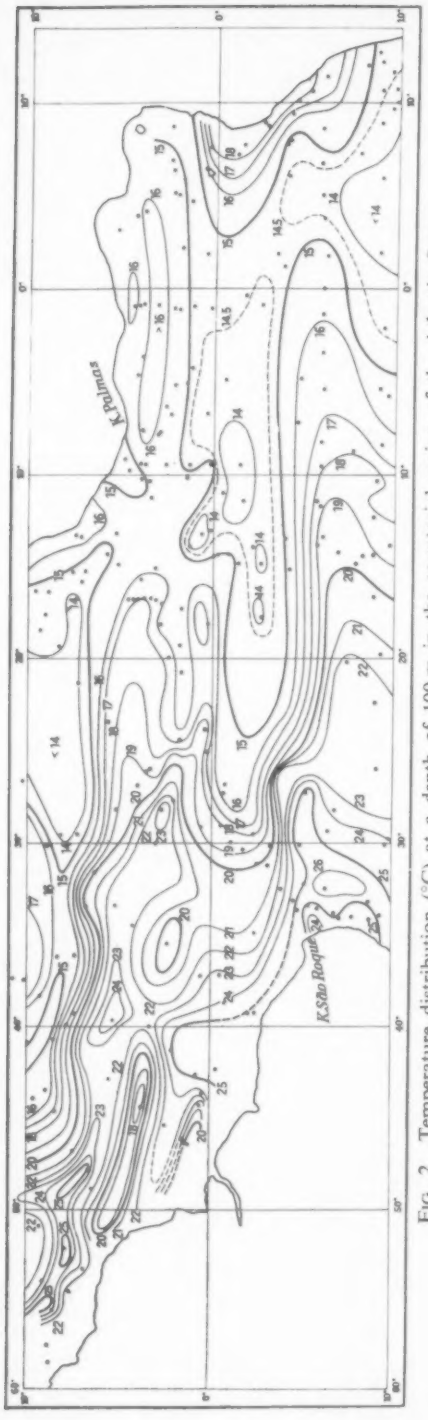

FIG. 2. Temperature distribution (°C) at a depth of 100 m in the equatorial region of the Atlantic Ocean.

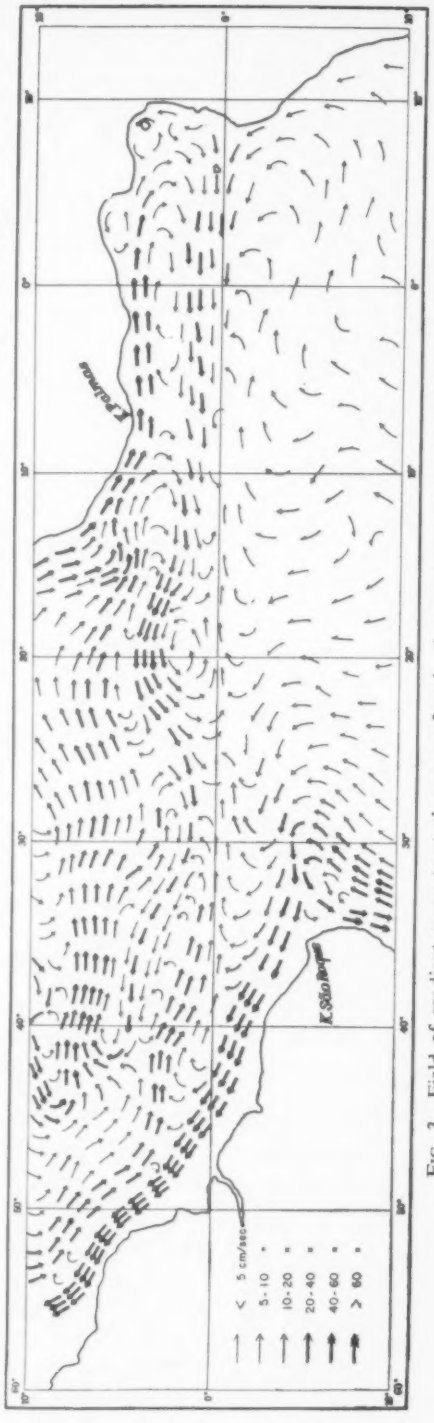

FIG. 3. Field of gradient currents at the sea surface in the equatorial region of the Atlantic Ocean.

Vol 6 1959

1947), the Equatorial Undercurrent seems to be the direct consequence of the *zonal* slope of isobaric surfaces at the equator.

If a simple model is assumed, such that under stationary conditions the zonal pressure gradient at the equator is balanced by frictional forces only, including the wind stress at the sea surface, an estimate of the velocity of the undercurrent is easily obtained. The slope of isobaric surfaces, γ in the western Atlantic equatorial region, between stations *Atlantis* 1180 ($\phi = 0° 24'$N, $\lambda = 43° 33'$W) and *Meteor* 212 ($\phi = 0° 36'$N, $\lambda = 29° 12'$W) is given in Table 1.

Table 1.

Isob. surf. (dbar)	0		50		100	150	200
$\gamma \times 10^8$ (obs.)	4·3		3·3		1·4	~0·4	0
$\alpha z \times 10^8$	4·0		3·0		2·0	1·0	0
z (*m*)	200	175	150	125	100	50	0
$u(z)$ cm sec^{-1}	− 26·0	− 12·1	− 2·4	3·6	6·5	5·7	0

The slope γ is maintained by the wind stress, τ, exerted by the Trade Winds. With an unaccelerated flow at the equator, the equation of motion reduces to

$$\mu \frac{d^2 u}{dz^2} = g\,\gamma \tag{1}$$

where μ (cm^2 sec^{-1}) is the kinematic eddy viscosity coefficient, u the current velocity to the east, and z is measured positively upward from the bottom of the current, such that $z = h$ is the depth of the whole current. At the bottom, $z = 0$, the velocity $u = 0$, and the condition

$$\int_0^h u dz = 0$$

prescribes that there is a net balance of zero, under stationary conditions, for the total flow in the zonal direction under the equator.

As shown in Table 1, γ is a function of z. It is assumed that $\gamma = 0$ at $z = 0$, at a depth of 200 metres. The observed $\gamma(z)$ may be approximated by $\gamma = \alpha$, with $\alpha = -2 \cdot 10^{-12}$ cm^{-1} for a rough estimate.

With these conditions, it follows from (1) that

$$u(z) = \frac{\alpha g}{6\mu}(z^3 - \tfrac{1}{2} z h^2) \tag{2}$$

This parabolic velocity profile fulfils the boundary condition $u = 0$ at $z = 0$ and shows that $u = 0$ also at $z = h/\sqrt{2} = 0{\cdot}71\,h$. At this depth there is a current reversal. The surface waters move to the west, driven by the east wind component. At a depth of $0{\cdot}71\,h$ the water starts to flow in the opposite direction, toward the east. With $\alpha = -2 \cdot 10^{-12}$, $h = 200$ m and $\mu = 50$ cm^2 sec^{-1}, the vertical velocity distribution $u(z)$ is given in Table 1.

Since the slope of the sea surface can be expressed in terms of the wind stress, τ, at the sea surface, it follows from (2) that

$$\tau/\rho = \mu\,(du/dz)\,h = \tfrac{5}{12}\,\alpha\, g h^2$$

With the same numerical values for α and H, it follows that the wind stress which maintains the slope is $\tau = 0{\cdot}33$ dynes/cm^2.

If the stress of the trade winds at the surface vanishes, and the assumption is made that the slope is maintained for some time, the water in the equatorial region would flow 'downslope' from west to east between the surface and the bottom. The surface velocity for the east-going current would be $u_0 = -\alpha gh^3/3\mu = 106$ cm/sec This is a maximum estimate, since with a vanishing wind stress and eastward motion throughout the whole layer, the slope of the sea surface would start to diminish rapidly.

This simple model is not intended to explain any details of the dynamics of the Undercurrent. It only shows that the magnitude of the forces that are involved is reasonable and that pressure gradient and friction do not contradict each other. The eastward slope of the upper isobaric surfaces in the vicinity of the Atlantic equator seems strong enough to provide the necessary hydraulic head for a narrow east going current between about 2° N and 2 °S. Under normal conditions when drift currents produced by the Trades are superimposed, the eastward slope current will not prevail at the surface. It may, however, be found at subsurface levels.

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Vol 6 1959

Micro-organisms as indicators of hydrological phenomena in seas and oceans—III

Distribution of water masses in the central part of the Pacific Ocean (according to microbiological data)

A. E. KRISS, S. S. ABYZOV and I. N. MITZKEVICH

Abstract—(1) A study of the vertical distribution of the numbers of heterotrophs was carried out at 56 stations in the Pacific Ocean along two parallel sections (174 °W and 172 °E from 37 °N to 41 °S). The equatorial-tropical zone is distinguished by a relative abundance of heterotrophs assimilating only easily accessible organic matter.

(2) The equatorial-tropical waters are found at various subsurface levels in the sub-tropical areas. A very thick layer was observed in the northern hemisphere along 172 °E where it was carried northwards by the Kuroshio current. Some strata of equatorial-tropical waters were also detected at considerable depths (6000–7000 m, 7500–9000 m) in the Kermadec trench.

(3) In the equatorial zone two layers of water, 375–550 m and 700–1800 m, were similar in their organic content to Arctic and Antarctic waters, these layers expanded considerably north and south from the equatorial zone. This circulation of waters from high latitudes across the equator also occurs in much deeper layers, judging from a layer of water with few heterotrophs observed near the equator at depths of 2500–4500 m.

ONLY one report on the microbiology of the Central Pacific Ocean is known; other investigations were carried out mostly in the northwestern part and near the Californian coast. The latter (MORITA and ZOBELL, 1955) deals only with the microbial bottom population.

During the *Vitjaz* expedition in 1957–58 63 microbiological stations were made along two parallel sections from 37 °N to 41 °S (Fig. 1). Twenty-seven of these with little deviation were along 174 °W and thirty-six along 172 °E. Water samples were taken at depths of 0 m, 10 m, 25 m, 50 m, 75 m, 100 m, 150 m, 200 m, 250 m, 300 m, 400 m, 500 m, 600 m, 800 m, 1000 m, 1500 m, 2000 m, 2500 m, 3000 m, 4000 m, 5000 m, 6000 m, 7000 m, 8000 m, 9000 m, 10,000 m and sometimes from intermediate layers. At 25 stations the entire water mass from surface to bottom was sampled and at 31 stations it was investigated only to 2000 m, 2250 m or 2500 m. A total of 1039 samples were analysed. At 7 stations only mud samples were taken.

Water samples were taken with all microbiological precautions in the *Vitjaz* laboratory. Samples were filtered immediately through the membrane ultra-filters N 2 in amounts of 50 ml. Then the filters were put with their rear (back) side on the surface of meat peptone agar prepared with Pacific Ocean water in the Petri dishes to incubate for 3-4 days at 25–30 °C. The colonies were then counted and representatives of the different colonies separated and placed on agar slants made with the same formula. The collection contained about 2000 strains.

The curves of Figs. 2, 3 and 4 show the vertical distribution of heterotrophs (grown on protein media) in Pacific Ocean water with the stations grouped for comparison according to geographic zones. With the two meridional sections, the northern

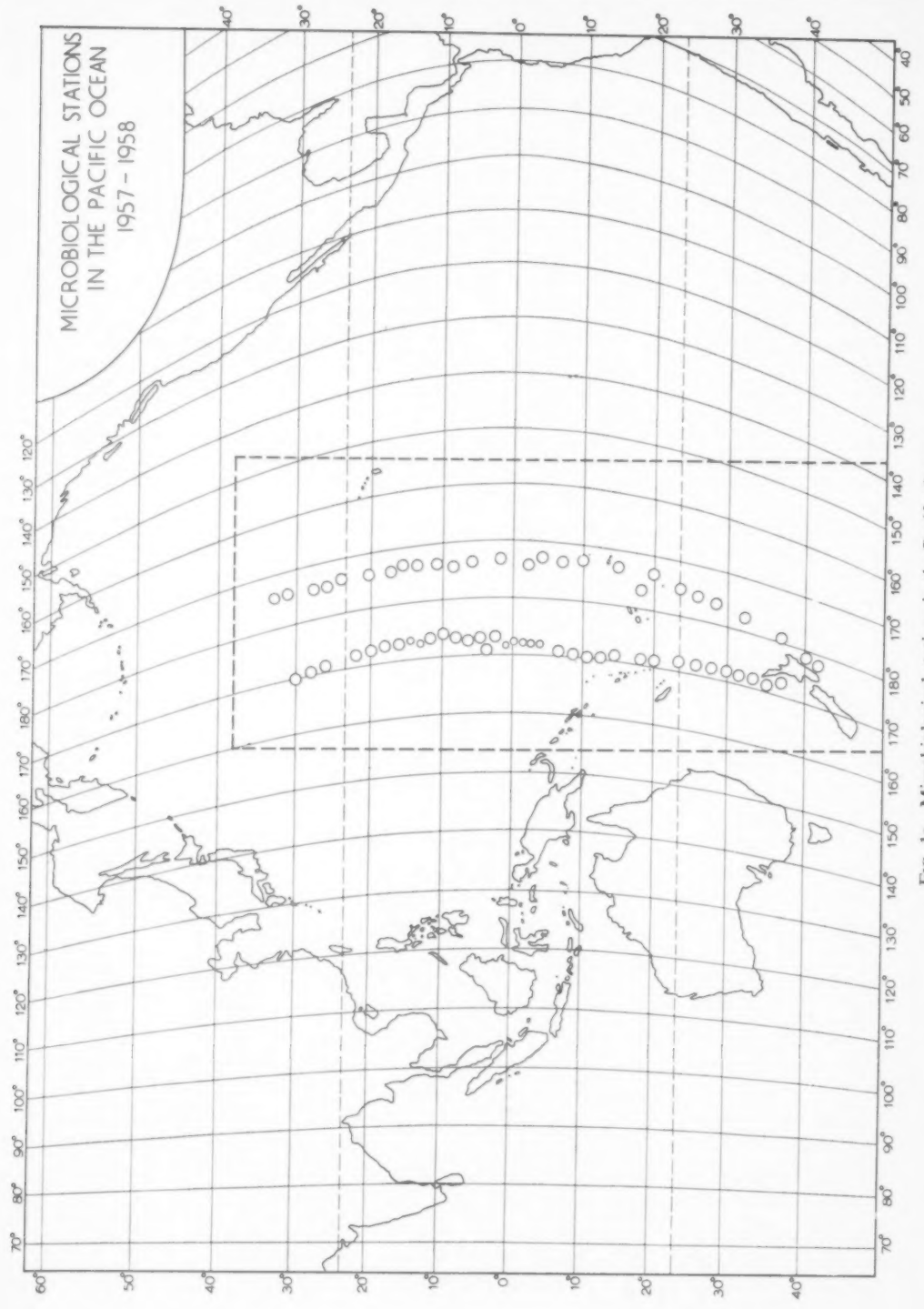

FIG. 1a. Microbiological stations in the Pacific Ocean.

Vol
6
1959

and southern subtropical, tropical and equatorial zones in the central Pacific, were crossed twice.

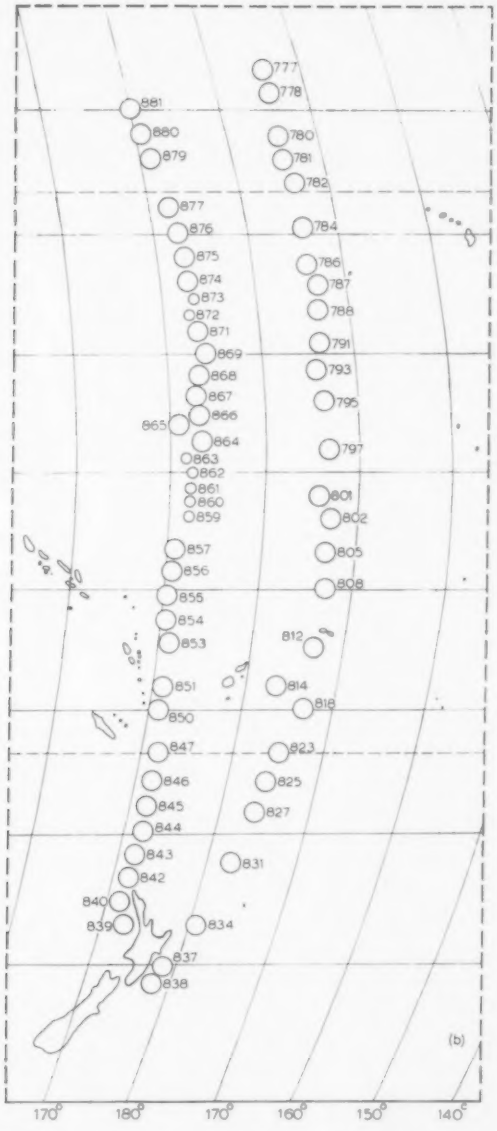

FIG. 1b. Microbiological stations in the central regions in the Pacific Ocean.

The concentrations of micro-organisms utilizing easily assimilated organic matter were quite different along the two sections in the subtropical area of the Northern hemisphere. The stations on the eastern section (174 °W) had a very low concentration of heterotrophs (Fig. 2). In most cases either a single colony only (from a 50 ml

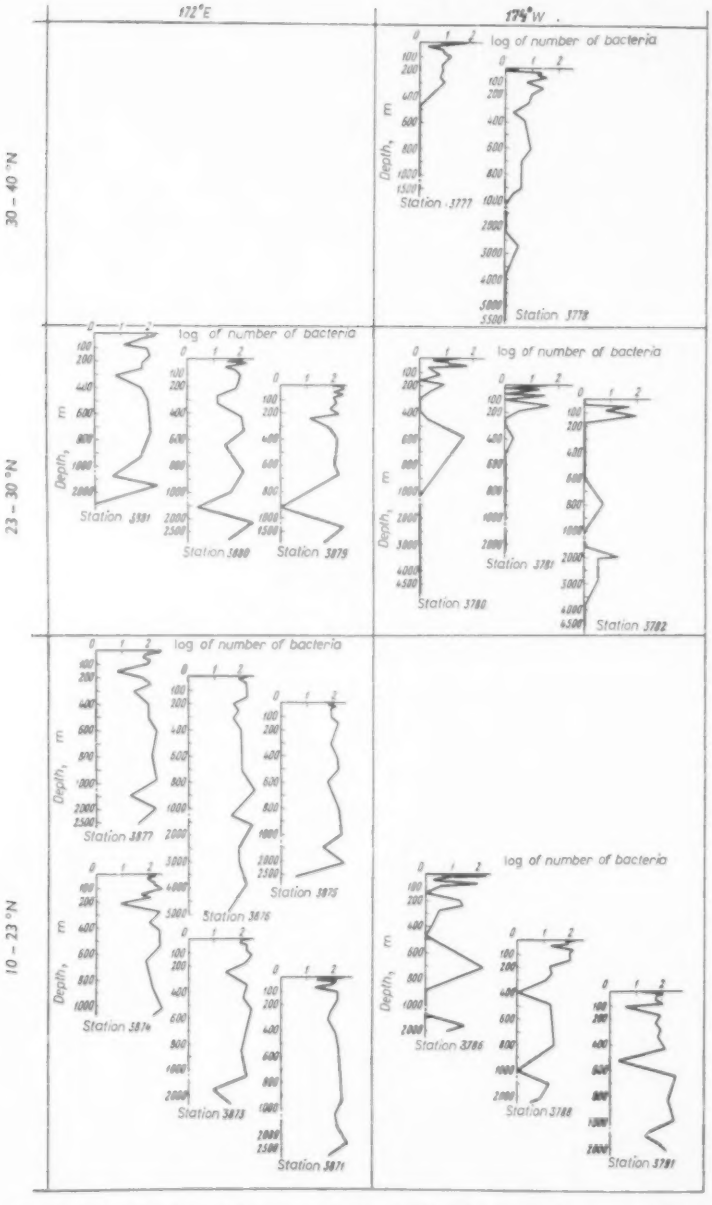

FIG. 2. Vertical distribution of heterotrophs in the Pacific Ocean between 40 °N and 10 °N (number of bacteria per 50 ml volume of water).

Vol 6 1959

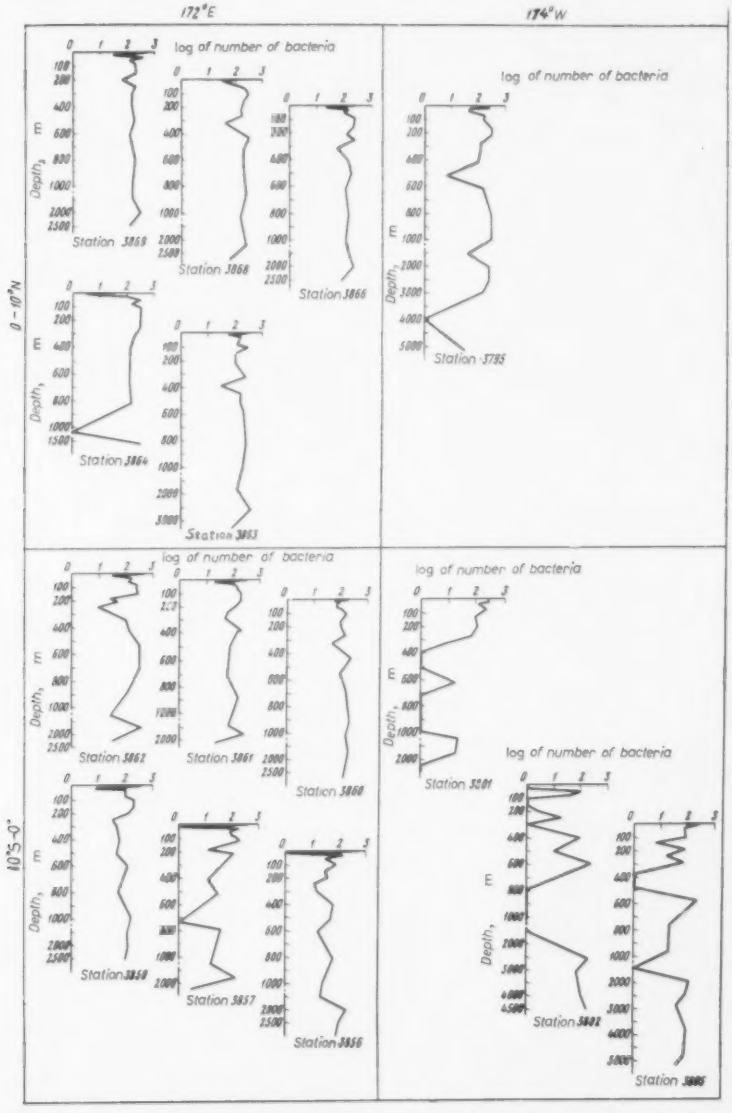

FIG. 3. Vertical distribution of heterotrophs in the Pacific Ocean between 10 °N and 10 °S (number of bacteria per 50 ml volume of water).

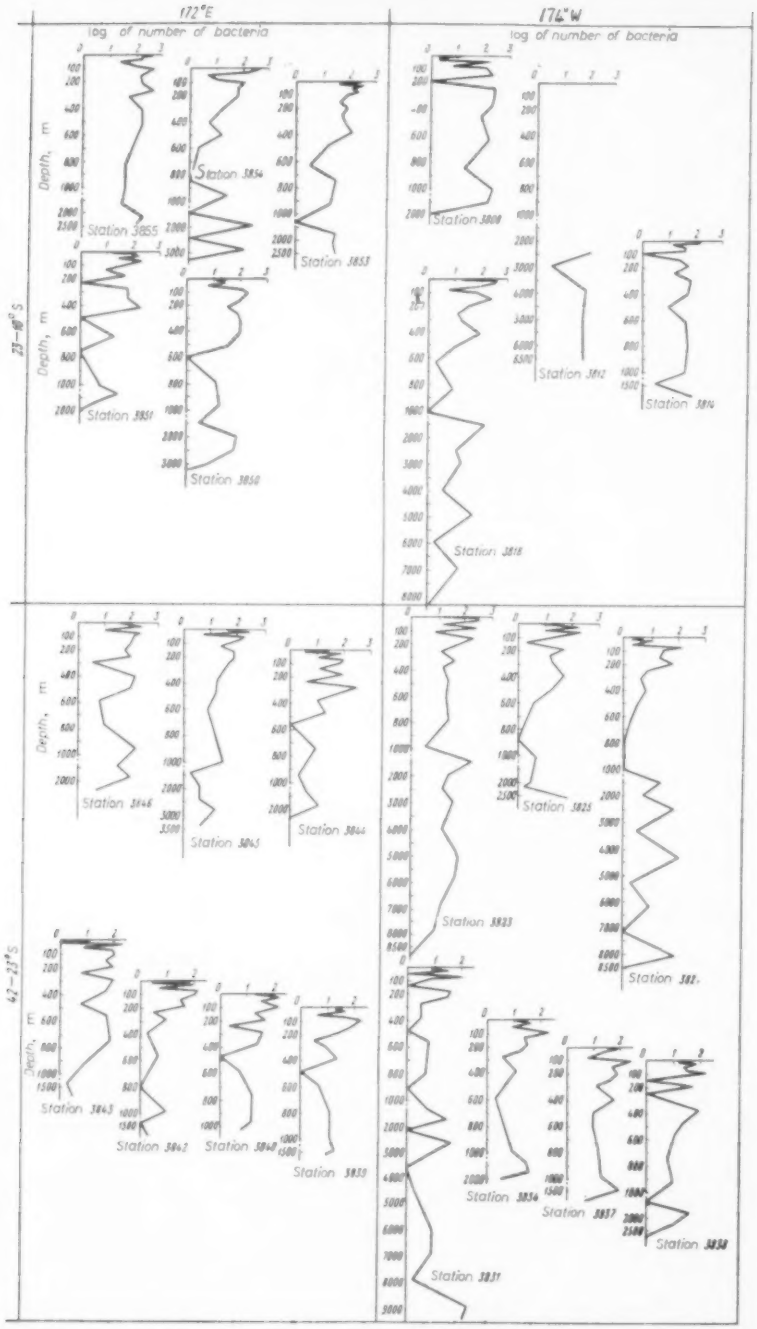

FIG. 4. Vertical distribution of heterotrophs in the Pacific Ocean between 10 °S and 42 °S (number of bacteria per 50 ml volume of water).

Vol 6 1959

sample) grew or the filter surface was sterile. Somewhat higher numbers of heteorotrophs occurred only in some layers.

In contrast on the section along 172 °E the same quantity of water yielded in almost all cases tens and hundreds of colonies. One concludes that a considerable concentration of easily assimilated organic matter was present in the subtropical water. Apparently the Kuroshio current, a ramification of the northern equatorial current, transports equatorial water enriched by organic matter to these regions. Beyond the influence of the Kuroshio the number of heterotrophs in the water mass was approximately the same as in the southern half of the Sea of Okhotsk (KRISS *et al.*, 1952) and the adjoining area of the Pacific Ocean (KRISS and BIRJUZOVA, 1955).

In the tropical zone north of the equator the differences between the west and east sections were somewhat less pronounced (Fig. 2). Along 174 °W the concentration of heterotrophs at many depths increased by tens and hundreds per 50 ml sample as in the western section, but there were a great number of layers with relatively few heterotrophs.

The same phenomenon also was observed in the equatorial area (10 °N–10 °S), a zone relatively rich in organic matter easily assimilated by the micro-organisms. With few exceptions almost the entire equatorial water mass contained considerable numbers of heterotrophs which grew on protein media (Fig. 3). In the northern half of the equatorial zone especially, hundreds of colonies developed on the filters of 50 ml samples. Along 172 °E, on the other hand, the numbers of heterotrophs do not fluctuate as sharply from one subsurface level to another as in the eastern section.

In tropical and subtropical areas of the southern Pacific Ocean a gradual decrease in the numbers of heterotrophs was observed toward the south, i.e., from tens and hundreds per 50 ml of water to fewer than ten (Fig. 4). At stations in the immediate vicinity of New Zealand, however, there was some increase in the number of bacteria.

In contrast to this process of the impoverishment in unstable forms of organic matter in Pacific Ocean water, the layers derived from equatorial-tropical waters were rich in heterotrophs as was observed when approaching the subantarctic zone. The hydrological structure of the Central Pacific Ocean is more complex in the southern hemisphere than in the northern. No sharp differences, however, were observed there between the eastern and western sections.

The microbiological data makes it possible to ascertain the hydrological structure in the Central Pacific Ocean. The quantitative distribution of heterotrophs indicates water with varying concentrations of readily assimilated organic matter. Differences in the origin of these waters cause the fluctuations in concentration. Equatorial-tropical water generally is characterized by relatively high amounts of easily assimilated organic matter (KRISS *et al.*, 1959), while in higher latitudes the water is deficient in these forms of organic matter and consequently there are fewer heterotrophs. The distribution of water with high and low concentrations of heterotrophs at all stations along the sections in the central part of the Pacific Ocean is shown in Fig. 5 (a). Along 172 °E, as along 97–99 °E in the eastern Indian Ocean, the water is rich in non-humus organic matter in the equatorial-tropical area where almost the entire water column contains relatively great numbers of heterotrophs. Apparently, this is due to the influence of the Pacific Ocean Zone rich in life forms and called the *Coral Sea* in the broad meaning of the word. In the subtropical areas the equatorial-tropical waters were found at varying depths. Along 172 °E in the southern hemisphere,

(except stations in the vicinity of New Zealand) this water occurred at depths of 25 m, 75–300 m, 250–550 m, 500–1200 m, 1000–3000 m. In the northern hemisphere, however, it formed a stronger current flowing toward high latitudes.

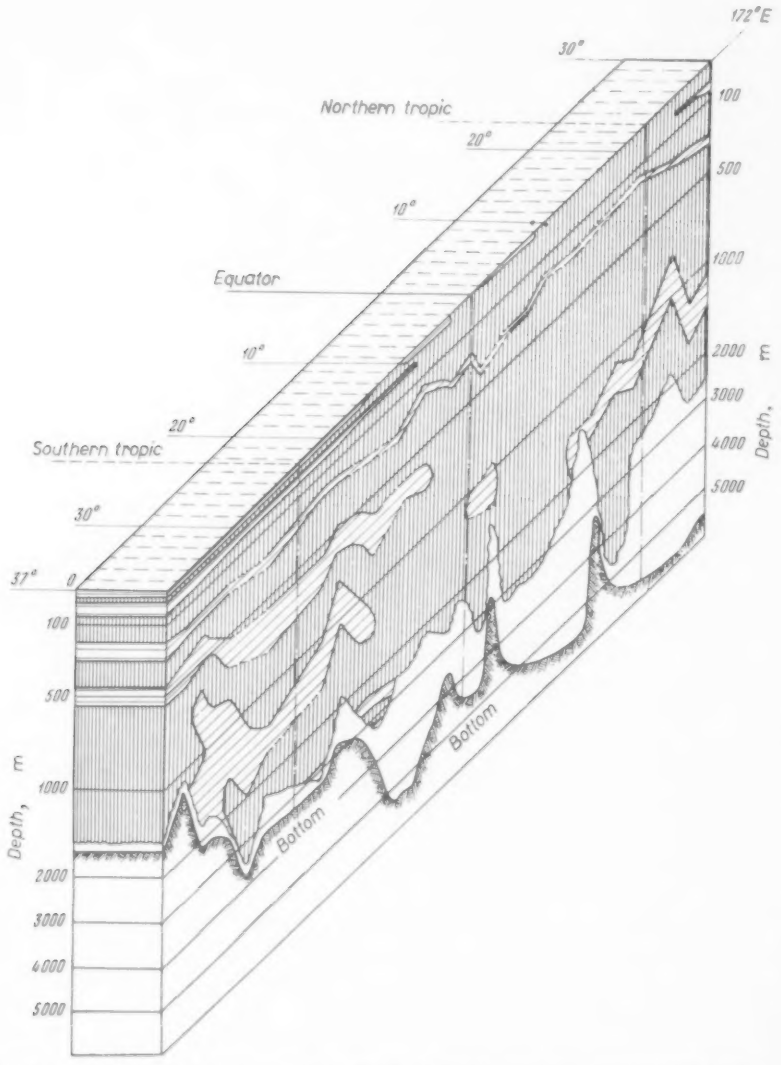

FIG. 5 (a). Hydrological structure of the Pacific Ocean (based on microbiological data).

The circulation from high to low latitudes was observed at various levels of the water column. Thus, water with small concentrations of readily assimilated organic matter was obtained at depths of 30–70 m, 150–400 m, 350–800 m, 650–1900 m, in the subtropical and tropical zones of the Pacific Ocean. Of particular interest are the water layers in Fig. 5 (a) connected by the dotted lines at depths of 150–200 m, 200–250 m, 250–300 m, 350–400 m, with few heterotrophs, identified at many localities

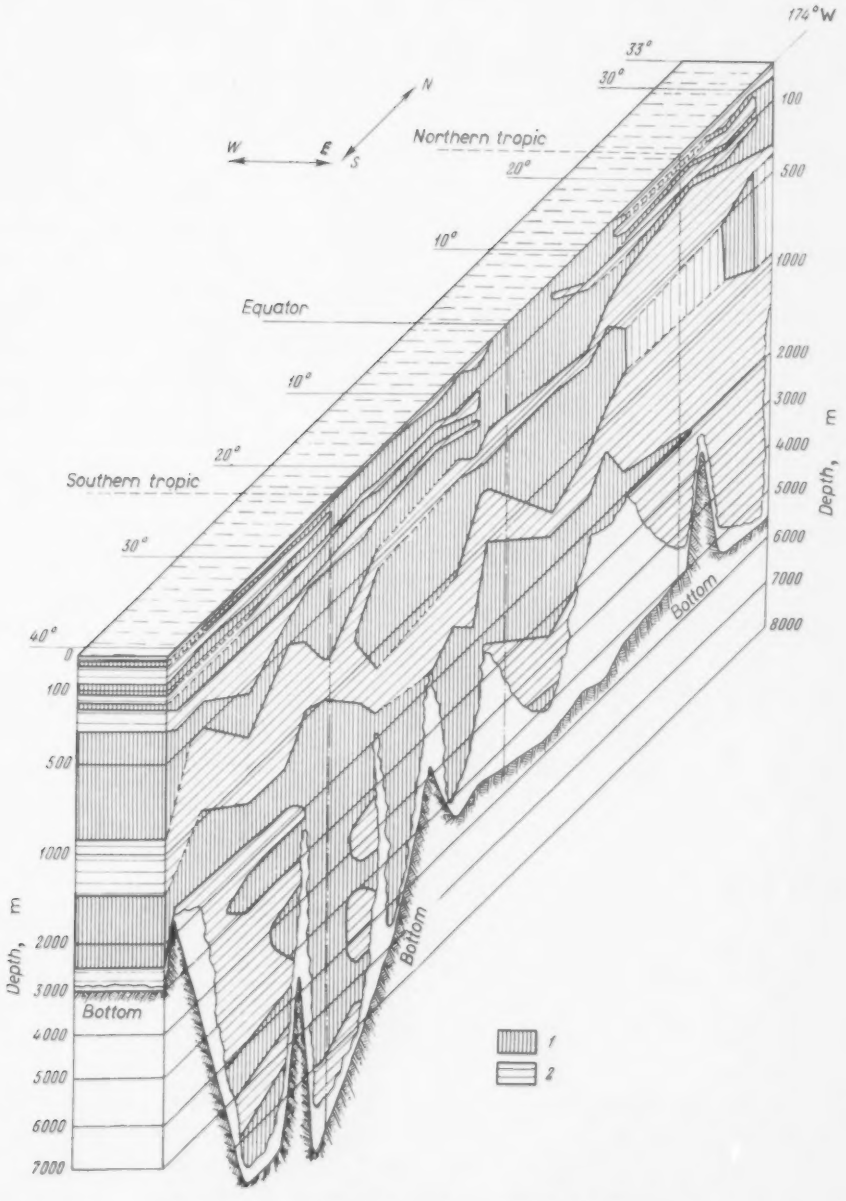

FIG. 5 (b). Hydrological structure of the Pacific Ocean.
1 – Layers of water of equatorial-tropical origin with high concentrations of bacteria.
2 – Layers of water of Arctic–Antarctic origin with low concentrations of bacteria.

along 172 °E from 30 °N to 37 °S. There is, however, no certainty that these actually represent a single water layer extending from high latitudes to equatorial-tropical area along 172 °E.

East of the 180th meridian, along 174 °W, new details of the hydrological structure of the Central Pacific Ocean Fig. 5 (b) were revealed. North of 10 °N the equatorial-tropical water was confined to comparatively narrow streaks, chiefly in the upper layer. The main water mass in the subtropical and tropical areas of the northern hemisphere is of other origin, apparently related to Arctic water poor in readily assimilated organic matter.

South of 10 °S, along 174 °W the equatorial-tropical water flows at depths of 25 m, 60–100 m, 100–250 m, 225–800 m, 500–1200 m, 950–1800 m, 2200–3000 m, 3500–5200 m and fills the Tonga Trench. Some streaks of this water were also detected at considerable depths (6000–7000 m, 7500–9000 m) in the subtropical area in the Kermadec trench.

Along 174 °W compared with the section along 172 °E there is a stronger invasion of Antarctic water as shown by a greater number of, and by thicker water layers, with few heterotrophs. Of particular interest and great significance is the evident penetration into the equatorial Pacific Ocean of water with a content of organic matter similar to that of the Arctic and the Antarctic. This water Fig. 5 (b) forms two almost uninterrupted layers in the equatorial zone (10° N–10° S) at depths of (1st layer) 450–550 m, 400–550 m, 375 m, 450–550 m, 400–500 m; (2nd layer) 1200–1600 m, 1400–1650 m 1000–1400 m, 700–1000 m, 1000–1800 m. North and south from the

Table 1. *Distribution of heterotrophic micro-organisms on various latitudes in the Eastern and Western halves of the Pacific Ocean*

Latitudes	*Longitude* – 172 °E					*Longitude* – 174 °W				
	Number of samples investigated	*Number of colonies on filters*				*Number of samples investigated*	*Number of colonies on filters*			
		0	*below* 10	*below* 100	*over* 100		0	*below* 10	*below* 100	*over* 100
40–23 °N	54	0	9·3	44·4	46·3	98	26·5	53·0	20·5	0
23–10 °N	111	0	3·6	37·8	58·6	54	7·4	25·9	44·5	22·2
10 °N–10 °S	196	1·0	3·1	32·1	63·8	82	17·0	12·2	30·5	40·3
10–23 °S	96	7·3	15·6	49·0	28·1	64	4·7	21·9	48·4	25·0
23–42 °S	123	1·6	32·6	48·8	17·0	149	6·0	34·9	48·3	10·8

The figures express the relationship between the water samples with the corresponding number of colonies and the total number of the samples studied, %.

equatorial zone these layers become considerably wider. Until the habitats of the particular species of heterotrophs in the Pacific Ocean become known, it is difficult to say whether waters crossing the equator are of Arctic or Antarctic origin. It is probably that in the equatorial zone the water layers low in easily assimilated organic substances represent waters from the convergence of waters flowing from the Arctic and Antarctic. At any rate the microbiological data confirm the supposition of an exchange of water masses in the Pacific Ocean across the equator in deep layers of the Central and Equatorial Waters.

Vol 6 1959

In the scheme proposed by SVERDRUP *et al.*, (1946), in the subtropical areas of the Pacific Ocean the Central Water (approximately 200–700 m) lies below the surface water (approximately 0–200 m) but above Intermediate Water (approximately 700–2000 m). Still deeper, the Antarctic Deep Water and similarly in the northern hemisphere – Arctic Deep Water flows above the Bottom Water. These authors state that in contrast to the Atlantic, in the depths of the Pacific Ocean little or no transport of waters across the equator is to be observed.

It is now evident that the organic composition of water (heterotrophs being indicators) may serve as a reliable indicator to determine the origin of the water masses. These indicators have helped to discover in the Pacific Ocean the circulation of water across the equator in layers which, according to SVERDRUP *et al.*, (1946), topographically correspond to the deep layer of the Equatorial Water and to the Intermediate Water layers which obviously are dissimilar in their organic content. In view of the fact the samples of water at many microbiological stations were collected no deeper than 2000–2500 m it is difficult to say whether or not an exchange of waters across the equator between the north and south Pacific Ocean occurs in the deeper layers and also at the bottom as in the Atlantic Ocean (DEFANT, 1957). Indirect observations (discovery of a water layer with few heterotrophs at the depth of 2500–4500 m) near the equator suggests the possibility that waters from the high latitudes enter the equatorial zone also at great depths.

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The content of heavy oxygen isotope in the water masses of the Philippine Trench

W. DANSGAARD

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Abstract—The H_2O^{18} content of a group of 11 water samples from 7558 to 9864 m depth was compared with that of another group of 4 samples from 3830 to 4202 m depth. The difference was 0·1 ± 0·2 p.p.m., i.e. not significant. Approx. 100 p.p.m. should be expected in case of equilibrium.

INTRODUCTION

IN natural oxygen the heavy isotope O^{18} occurs in quantities approximately equal to ·20 atom % or 2000 p.p.m., depending somewhat on the origin of the oxygen.

Except for surface water the H_2O^{18} content in ocean water, unmixed with fresh water, is rather constant (± 0·5‰ of the mean value) from place to place (EPSTEIN and MAYEDA, 1953). By accurate measurements on samples from the Atlantic and Pacific oceans EPSTEIN and MAYEDA (1953) found that for a given salinity the deeper samples are poorer in O^{18}. The deepest sample investigated was from a depth of 7500 m at 00° 13′S, 18° 26′W. It had a salinity close to the mean for ocean water but a relatively low H_2O^{18} content. They suggested that, 'a possible explanation of this fact is that this sample was collected from one of the deep ocean currents. These currents resulting from sinking polar waters, flow to the equatorial region and thus add what was originally surface water. Such surface water could result from the addition of melted snow at first and then the subsequent freezing out of fresh water from the less saline water of low O^{18} content, the latter process being accompanied by little isotopic fractionation.'

For evident reasons a possible gravitational fractionation should be looked for in stagnant water at great depths. Under such circumstances, sedimentation of H_2O^{18} due to gravity should result in an exponential increase in H_2O^{18} content downwards, analogous to the situation for the components in the atmosphere (HAGELBERGER *et al.*, 1951) and in solutions. In Lake Baikal considerably heavier water was found at 1600 m than at 600 m depth (VERESCAGIN *et al.*, 1934). This was ascribed to differences in H_2O^{18} content (MENDELEJEV, 1935).

COLLECTION OF DEEP SEA SAMPLES

The samples investigated in this work were collected by the *Galathea* Expedition 1950–52, the Danish Deep-Sea Expedition Around the World.

Unfortunately, the *Galathea* Expedition took no samples for isotope analysis at mid-depths in the Philippine trench. However, knowledge of the ocean currents is sufficient to make possible the use of samples taken elsewhere as representative of the water at 4000 m in the Philippine Trench. The easterly current from the Atlantic south of Africa sends off a branch to the deep water of the Indian Ocean, although the main current flows south of Australia. There it turns north toward the Philippines.

Vol 6 1959

Table 1

Sample No.	*Station*	*Position*	*Depth* (m)	H_2O^{18} *diff. from No.* 10 p.p.m.	*Temp.* °C	*S.* ‰	$(\zeta_t - 1) \cdot 10^3$ g/cm³	0° ml/l.
1	232	9° 03′S 49° 22′E	4070	÷ 0·4	1·34	34·70	27·81	4·08
2	236	3° 22′S 45° 50′E	3830	+ 0·1	1·87	·78	·83	3·40
3	236		4070	+ 0·1	1·80	·76	·82	3·52
4	407	12° 10′N 114° 56′E	4202	0·0	2·48	·61	·65	2·40
		Mean		*÷ 0·05 ± 0·2*				
5	412	11° 13′N 126° 21′E	8595	+ 0·2				
6	431	10° 21′N 126° 38′E	7558	+ 0·4	2·01*	34·69	27·75	3·04
7	431		8503	÷ 0·5	2·27	·69	·73	3·11
8	431		9165	0·0	2·35	·67	·72	2·88
9	431		9636	+ 0·5	2·50	·67	·71	3·27
	433	9° 51′N 126° 51′E	3797		1·57	·67	·77	3·27
	433		4252		1·57*	·69	·78	3·15
	433		4724		1·70	·67	·77	3·11
	433		5669		1·80	·69	·77	3·12
	433		6614		1·95	·69	·76	3·36
10	433		7511	*0·0*	1·97*	·68	·75	3·20
11	433		8456	÷ 0·3	2·18	·66	·72	3·16
12	433		9498	+ 0·1	2·28*	·66	·71	2·93
13	433		9589	÷ 0·1	2·47	·69	·72	3·31
	440	10° 25′N 126° 40′E	3068		1·61	·65	·75	2·92
	440		5900		1·83	·67	·75	3·01
	440		7788		2·13	·68	·74	2·90
14	440		9770	+ 0·1	2·56	·67	·70	3·19
15	440		9864	+ 0·3	2·55	·67	·70	3·11
		Mean		*+ 0·06 ± 0·05*				

*Thermometer not absolutely reliable.

From Bruun and Kiilerich (1955).

Salinity and temperature indicate that there is no significant addition of fresh or Polar water on the way. Therefore, samples 1–3 taken in the Indian Ocean and 4 taken in the China Sea have a chemical and isotopic composition close to that of the water at 4000 m depth in and around the Philippine trench.

PREPARATION AND MEASUREMENT

The water samples were prepared for mass spectrometric O^{18} determination by equilibration with CO_2. When the exchange of oxygen atoms had come to an equilibrium, the CO_2 was extracted and used for the isotope measurement.

The differences in H_2O^{18} content between the CO_2 samples were thus equal to those between the corresponding water samples, the fractionation factor in the equilibrium process being negligible here.

CO_2 in equilibrium with one of the water samples (No. 10 in Table 1) was used as a reference. The fifth column (Table 1) shows the differences measured with a modified Consolidated mass spectrometer.

The samples were divided into two groups :

1. Four samples taken at depths of approx. 4000 m.
2. Eleven samples taken at depths from 8000 to 10,000 m.

The difference between the mean values for the two groups, $0{\cdot}1 \pm 0{\cdot}2$ p.p.m. is not significant.

DISCUSSION

The degree of isotopic fractionation to be expected in water stagnant over a very long period of time is calculated below.

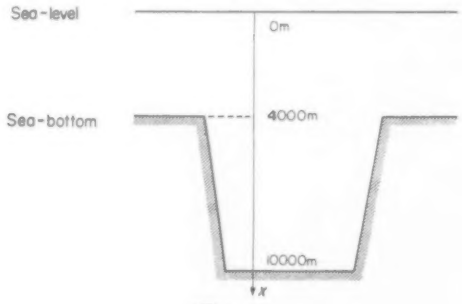

FIG. 1.

A schematic drawing is given of a trench (Fig. 1), the bottom of which is 6000 m below the surrounding sea bottom, which is in turn 4000 m below the sea level. The depth, x, is measured along an x – axis as shown in the figure. The two groups of samples mentioned above were taken at $x = 4000$ m and $x = 8000$ to 10,000 m, respectively.

The equation giving the H_2O^{18} content, a, as a function of x, when the isotopic sedimentation has come to equilibrium, is then

$$a_x = a_{x0} \exp\left(\frac{M - M'}{RT}\, g\,(x - x_0)\right), \qquad (1)$$

Vol 6 1959

a_{x0} being the value for a when $x = x_0$, (approx. 2000 p.p.m.), M and M' the molecular weights of H_2O^{18} and H_2O^{16} (20 and 18), g the gravity (980 cm/sec^2), R the gas constant ($8{\cdot}3{\cdot}10^7$ ergs/mol. °C) and T the absolute temperature (275 °K). Since the water down to $x = 4000$ m is certainly well mixed, a possible fractionation should be looked for in the trench as such, i.e. from $x = 4000$ m to $x = 10{,}000$ m, and the formula comes out as

$$a_x = a_{4000} \exp\left(\frac{M - M'}{RT}\, g\,(x - 4000) \cdot 100\right).$$

This formula gives a value for a at the bottom of the trench

$$a_{10000} = 1{\cdot}052 \cdot a_{4000}$$

and

$$a_{10000} - a_{4000} = 0{\cdot}052 \cdot a_{4000} \simeq 100 \text{ p.p.m.}$$

which is 500 times the accuracy of the experimental measurements.

The fact that no significant fractionation has been found indicates that the time for a complete exchange of the water masses in the trench is very short, relative to that needed for the sedimentation process to reach equilibrium. On the other hand this latter time is very long, probably of the order of magnitude of 10^8 years.

The uniformity in isotopic composition agrees with the observations of temperature, salinity and density made by VAN RIEL *et al.*, (1950, Snellius Expedition), BRUNEAU *et al.*, (1953, Swedish Deep-Sea Expedition) and BRUUN and KIILERICH (1955, *Galathea* Expedition). The last four columns of Table 1 show temperature, salinity, density and oxygen content at different depths (from BRUUN and KIILERICH, 1955).

MENDELEJEV (1935) reported a density difference of $3{\cdot}7 \times 10^{-6}$ g/cm^3 (after distillation) in samples from 600 and 1600 m in Lake Baikal, a difference ascribed to different H_2O^{18} contents. This corresponds to 33 p.p.m. With $x = 1600$ m and $x_0 = 600$ m equation (1) gives the highest possible difference due to gravitational fractionation:

$$a_{1600} - a_{600} = 1{\cdot}8 \text{ p.p.m.}$$

Therefore, if the reported density difference is due to different H_2O^{18} contents, the result cannot be ascribed to gravitational fractionation, not even if the water had had time enough to reach an equilibrium.

Acknowledgements—The author wishes to thank the *Galathea* Expedition for the samples, and especially the leader of the expedition, Dr. Phil. A. F. BRUUN, for his active interest in the work. Many thanks are also due to Profs. Dr. Phil. J. KOCH and Dr. Med. F. SCHØNHEYDER, and to Lektor Mag. Scient. A. KIILERICH for their never resting helpfulness.

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Vol
6
1959

The annual cycle of primary production in the Sargasso Sea off Bermuda

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Abstract—Gross and net primary production have been measured at bi-weekly intervals for 18 months in the North-western Sargasso Sea 15 miles SE. of Bermuda in 1500 fm of water. Ancillary data include temperature, salinity, phosphate, nitrite, nitrate, disolved oxygen, plant pigments, light penetration and incident radiation.

A seasonal cycle of production was observed with high levels in the winter and early spring, low levels in the later spring, summer and early fall. Gross production ranged from 0·13 to 2·00, averaging 0·44 g carbon assimilated/m²/day or 160 g C/m²/year. Net production ranged from 0·05 to 0·83 and averaged 0·20 g C/m²/day or 72 g C/m²/year.

Production was closely dependent upon vertical mixing, high levels occurring when the water was isothermal and mixed to or near the depth of the permanent thermocline (400 m), low levels being associated with the presence of a seasonal thermocline in the upper 100 m. Nutrient concentrations were extremely low (maxima of 1·8 μgAN/L and 0·16 μg AP/L as inorganic compounds in the upper 100 m) and showed little seasonal variability. Mixing and enrichment from the permanent thermocline is negligible, production being largely dependent upon the rapid re-cycling of nutrients within the upper 400 m.

The environmental factors controlling production which make this region differ from temperate or boreal waters, and which permit high production throughout the winter are: (1) low nutrient concentrations, (2) clear water, (3) relatively high incident radiation in winter, (4) a shallow winter mixed layer and (5) a rapid re-cycling of nutrients, possibly due to higher temperatures.

INTRODUCTION

IN THE fall of 1957 a study was initiated of the plankton biology and the related chemistry and hydrography at a single, fixed position in the North-western Sargasso Sea. This programme has been carried out from the Bermuda Biological Station, employing that Institution's research vessel *Panulirus*. The oceanographic station (referred to henceforth as Station 'S') is located 15 miles SE. of Bermuda (32°N, 65°W) in 1500 fm of water where conditions appear to be typical of the North-western Sargasso Sea; there have been no indications of island effects since the programme was started. The same position has been occupied since 1954 for the purpose of making routine standard hydrographic observations under a joint arrangement with the Woods Hole Oceanographic Institution. The biological programme may be considered as an adjunct to the continuing hydrographic studies.

The location of a truly oceanic station just 15 miles from land presents obvious advantages and unique features, since it may be occupied at frequent intervals throughout the year with a minimum expenditure of time and funds. Thus the study of annual cycles, seasonal and yearly variations, etc., of various oceanographic phenomena, never before possible, may be routinely undertaken. The present report will describe one aspect of the current programme, the seasonal cycle of primar production, as followed over a period of 18 months.

The annual rate of primary production in the open ocean has been a warmly debated subject in recent years. This has resulted primarily from the large discrepancy between the values reported for the Sargasso Sea by RILEY *et al.* (1949) and STEEMANN NIELSEN (1952). There have been various conflicting explanations for this lack of agreement (STEEMANN NIELSEN, 1952, 1954; RILEY, 1953; RYTHER, 1954, 1956a, 1956b) which have done little more than confuse the issue. More recently RILEY (1957) has attempted to resolve the problem by making new estimates for the same region by means independent of either of the earlier investigations. The three separate studies are summarized in Table 1 showing the methods employed and the results obtained in each case.

Table 1. *Summary of data for average daily gross primary production in the Sargasso Sea*, 1949–1957

Reference	*Method*	gC/m²/*day*
RILEY *et al.* (1949)	(oxygen change, light and dark bottles)	0·46
STEEMANN NEILSEN (1952)	C^{14}	0·05
RILEY (1957)	*in situ* 0_2 changes	0·27

Vol 6 1959

RILEY's latest figures were based upon *in situ* oxygen changes in the upper 300 m of water as estimated from physical oceanographic calculations of oxygen production and consumption. The present authors offer no criticism of this method, which has many advantages over the available experimental techniques. However, his oxygen data were obtained from four *Atlantis* cruises between 1932 and 1949, while his mixing coefficients were based on bathythermograph records taken in 1950-1952 by Weather Ship personnel at Station E. Hence, the resulting values of primary production can be no more than rough approximations. Furthermore, though data were obtained for more different times of the year than previously available, giving indications of considerable variability, no true picture of the seasonal cycle could yet be obtained.

It was obviously worthwhile, therefore, to undertake the study of primary oceanic production at the semi-tropical latitude of Bermuda on a seasonal and year-to-year basis. The objectives of this study were (1) to establish the annual rate of primary production, (2) to determine whether or not there is a well defined seasonal cycle similar to that observed in temperate and northern waters, (3) to study variations from year to year in the magnitude of production and characteristics of the annual cycle and (4) to study the chemical and physical factors which control and limit production in these waters.

METHODS

In the 18 month period between November, 1957 and April, 1959 Station S was occupied on more than 55 occasions for an average of better than three times per month. Although all operations were not carried out each time the station was occupied, a total of 34 complete hydrographic stations and 48 productivity measurements (each usually by two independent methods) were completed.

At each hydrographic station, measurements were made to an indicated depth of 2600 m of temperature, salinity, inorganic phosphorus, total phosphorus, nitrite, nitrate, silicate, and dissolved oxygen*. Two- to four-litre water samples were collected from five depths to 100 m for the analysis of plankton pigments by the method of RICHARDS with THOMPSON (1952) as modified by CREITZ and RICHARDS (1955). These samples were collected in two-litre, non-metallic samplers of the type described by VAN DORN (1956). In addition, a transparent and opaque 150 ml glass stoppered bottle was filled from each of these samples. To these bottles were added approximately 20 μ curies of $CO_3^{=}$. This was prepared essentially as described by STEEMANN NIELSEN (1952). For the first part of the study (November, 1957–April, 1958) the bottles were fastened to the hydrographic wire and lowered to the depths from which the samples were taken (usually 0, 10, 25, 50 and 100 m), where they were held for half the daylight portion of a day (i.e. noon to sunset). After April, the samples were collected from the respective depths to which 100 per cent, 50 per cent, 25 per cent, 10 per cent and 1 per cent of the surface light penetrated. The samples were then dispensed into bottles and C^{14} added as before, and the bottles were placed in a box cooled with running surface water and covered with a series of Kodak Wratten neutral density filters which transmitted the same fraction of the incident radiation as that to which the contained organisms had been exposed *in situ*. The samples were held in the filter box for 24 hrs. After exposure (by both the above mentioned methods) the contents of the bottles were filtered through HA millipore filters (pore size 0·5 μ), the filters were rinsed with 10 ml of 0·001 N HCl in a 3 per cent NaCl solution, dried in a desiccator for at least 24 hrs, and counted in a gas flow Geiger counter. The activity of the filtered material from the dark bottles was subtracted from that of the corresponding transparent bottles, and the net activity was used to calculate productivity by the method described by STEEMANN NIELSEN (1952) but with no correction for respiration. The resulting value is interpreted as net daily production (RYTHER, 1956b; STEEMANN NIELSEN and HANSEN, 1959).

The light penetration of the water was determined for each station with a Secchi disc and the extinction coefficient per metre was calculated according to the formula of POOLE and ATKINS (1929). This was checked periodically with a submarine photometer. Total incident radiation was recorded on shore in Bermuda with an Eppley 10 junction pyrheliometer. From the chlorophyll *a* concentration at the depths of penetration of 100 per cent, 50 per cent, 25 per cent, 10 per cent and 1 per cent of the surface light and the total daily incident radiation, daily gross production at each of these depths, was calculated by the method of RYTHER and YENTSCH (1957). Since pyrheliometer records were not available for the entire period, radiation values for this study were estimated from KIMBALL's (1928) tables.

The daily rates of net (C^{14} method) and gross (chlorophyll-light method) production for each of the five above mentioned depths was then plotted on graph paper as a function of depth and the areas of the two resulting curves were measured with a planimeter giving daily net and gross production beneath a square metre of surface. A sample graph showing these plots is given in Fig. 1.

*Salinity (Knudson method) and total phosphorus (Harvey method as modified by KETCHUM *et al.*, 1955) were determined at the Woods Hole Oceanographic Institution. All other analyses were completed by the senior author at the Bermuda Biological Station by the following methods : phosphate (MURPHY and RILEY, 1958), nitrite and nitrate (MULLIN and RILEY, 1955a), silicate (MULLIN and RILEY, 1955b) and oxygen (WINKLER method).

The accuracy of the 'chlorophyll-light' method in estimating gross primary production in continental shelf waters has been studied and is discussed by RYTHER and YENTSCH (1958). The method has not been similarly evaluated for the semi-tropical Sargasso Sea, and there appears to be no satisfactory way of doing so. It is based upon the following assumptions: (1) a constant assimilation number (photosynthesis per unit of chlorophyll at optimal light intensity), and (2) a constant and predictable relationship between photosynthesis at optimal light intensity and at any other naturally occurring intensity. The latter relationship which is used in the formulation of the method has, in fact, been found not to describe accurately the photosynthesis-light intensity curve of the Sargasso Sea plankton, in that it fails to account for light adaptation by the organisms (RYTHER and MENZEL, in press). On this account alone the resulting values are probably somewhat too low.

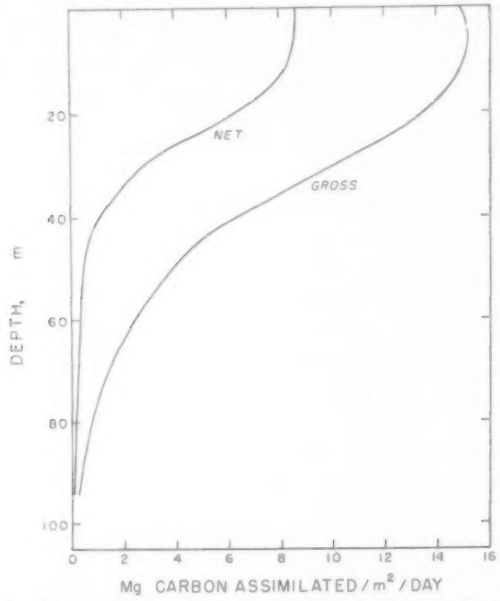

FIG. 1. Daily net and gross photosynthesis on January 30, 1959.

The assimilation number cannot be checked without an independent method of measuring total photosynthesis. C^{14} uptake (net photosynthesis) does not bear a constant relationship to total photosynthesis and hence cannot be used for this purpose. In relatively rich coastal waters the 'light and dark bottle' oxygen method may be used, but it is too insensitive for the oligotrophic Sargasso Sea. Therefore the assimilation number must be assumed to be the same as that of the continental shelf plankton. This assumption is partly justified by the fact that Sargasso Sea phyto-plankton populations are qualitatively and quantitatively similar to those of the off-shore continental shelf waters.

HYDROGRAPHY

The Sargasso Sea may be thought of as a lens of homogeneous water characterized by a salinity of almost exactly 36·50‰ and a temperature within a few tenths of a

degree of 18 °C. This lens tapers upward from a thickness of almost 500 m in the central Sargasso Sea to become nearly extinguished at its boundaries. In the vicinity of Bermuda it extends to a depth of some 400 m. Below this lens lies the permanent thermocline. Above it a seasonal thermocline extends to a depth of 100-150 m. At the latitudes of Bermuda and above, the seasonal thermocline may be destroyed in winter, the 18° water extending all the way to the surface. South of Bermuda, with less surface cooling, the upper thermocline persists throughout the year and merges more gradually with the 18° water below.

Two publications have recently appeared concerning this interesting '18° water.' The first by SCHROEDER *et al.* (1959) describes the remarkable climatic stability of this layer which, though seasonally in contact with the surface, has apparently retained its T-S characteristics for at least the 85 years for which records are available. The second, by WORTHINGTON (1959) is concerned with the geographical distribution of this water mass and, particularly, its seasonal formation at the surface at and North, of the latitude of Bermuda and its subsequent intrusion southward beneath the permanently stratified surface waters. Despite the mysterious retention of its temperature and salinity characteristics, the seasonal changes of the oxygen concentration of this water indicate that it is annually renewed, and this applies as well to the 18° water south of Bermuda where the surface waters never approach this temperature.

We are concerned here principally with hydrographic features around Bermuda and their relationship to primary production. Since our interests centre around the upper few hundred metres, the 18° water and its seasonal cycle of stabilization and overturn are especially pertinent. The paper by SCHROEDER *et al.* is based upon the hydrographic observations at Station S since 1954; hence we need not reproduce the record prior to 1957 here. However, their published data stop with June, 1958, and thus include only half the period in which we are interested. Therefore the temperature and salinity records to a depth of 900 m are presented here for the period November, 1957 through April, 1959 (Fig. 2).

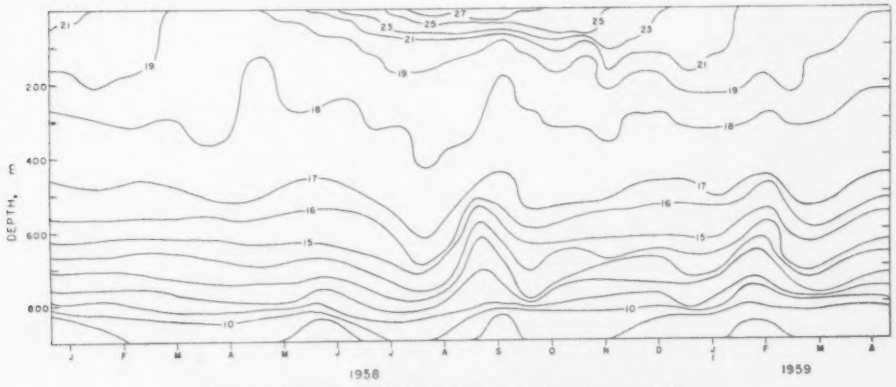

FIG. 2. The seasonal depth profile of temperature (°C).

The temperature record shows that the water was isothermal within a few tenths of 18° from the surface to almost 400 m from mid-February to mid-April, 1958. By April, the seasonal thermocline formed and was strongly developed throughout

the summer and early fall, extending some 50-100 m in depth. By mid-September, surface mixing and cooling became apparent, the mixed surface layer increasing in depth to about 200 m as the surface temperature dropped from 23° in November, to 19° in February, 1959. However, the surface water never cooled to 18° in 1959 and hence the water column did not become mixed to the depth of the permanent thermocline, as in 1958. The greater cooling and mixing in the early spring of 1958 may have been related to abnormally low air temperatures that year (Table 2). An

Table 2. Mean monthly air temperature at St. Georges, Bermuda, January–March

Month	1958	1959	*Mean* 1954–1959
January	64·0	64·0	64·3
February	61·6	65·0	64·0
March	63·8	66·5	66·7

examination of the thermal structure of the upper water since 1954 (SCHROEDER *et al.* 1959) reveals that 18° surface temperatures occur sporadically and probably not too frequently at the latitude of Bermuda. This hydrographic feature is believed to play an important part in the seasonal cycle of primary production and will be discussed later.

As described above, the 18° water is further characterized by a salinity of 36·50‰ and water of nearly this salinity thus occurs over most of the upper 400 m. However, at the very surface during that part of the year when the seasonal thermocline is present, the salinity is irregular and less predictable, but consistently lower than 36·50‰. This relatively fresher layer may be as much as 100 m in depth and at times (i.e. July and October, 1958) the surface salinity may drop to as low as 36·00‰ (Fig. 3). It is not inconceivable that this freshening may result from rainfall, for a

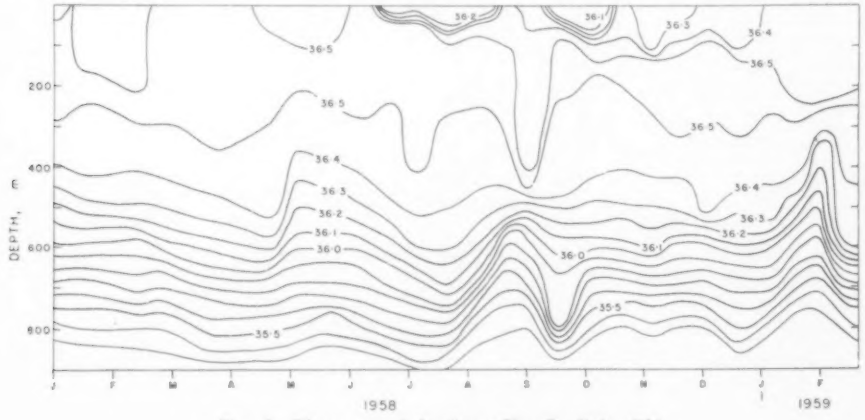

FIG. 3. The seasonal depth profile of salinity (‰).

total of 10-20 inches would dilute 100 m of water to the extent observed. W. H. Sutcliffe (personal communication) has found a rather close inverse correlation between rainfall and the mean salinity of the upper 100 m. The mean annual rainfall

at Bermuda is approximately 60 inches, and 5-10 inches per month is not uncommon. This does not take evaporation into consideration, which might be expected to exceed precipitation in the summer months, though ISELIN (1936) suggests that the higher summer humidity may actually retard the process. However, the explanation is almost incompatable with the observed constancy of the salinity of the 18° water. It is difficult to imagine this water warming and freshening simultaneously at the surface, but how or why it should return to exactly the same salinity when it cools to 18° is inconceivable.

The only other possible explanation for this freshwater surface layer is that it represents an intrusion of another water mass, presumably from the Gulf Stream. This has been seriously suggested but also presents difficulties. For one thing, there are no changes in other properties (temperature, chemical nutrients, etc.) accompanying the occurrences of low salinity surface water which one might expect with a different water mass. Furthermore, this phenomenon appears to be so widespread over much of the Western Sargasso Sea and such a regular feature throughout most of the late summer (ISELIN, 1936; FUGLISTER, 1947; SCHROEDER *et al.* 1959) that it is difficult to conceive of eddies or outflows of Gulf Stream water of such magnitude, persistence and regularity. Intriguing as this problem is, it need not concern us here, for the low salinity surface water appears to have no direct effect on the plankton biology of the region beyond contributing to the stability of the surface layer.

NUTRIENT CHEMISTRY

The seasonal cycle of the plant nutrients which have been studied (phosphate, nitrite, nitrate and silicate) may be treated rather briefly. Their concentrations in the euphotic zone, and even in the mixed layer in winter, are low and do not show the seasonal variations found in temperate and northern waters. The indications are

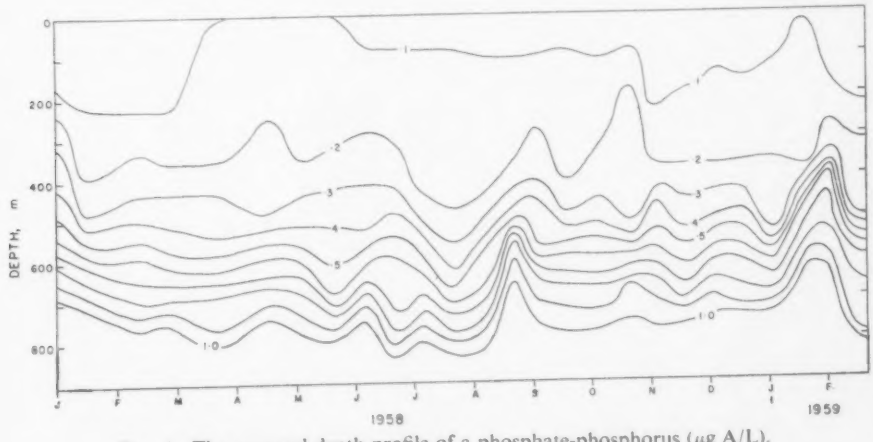

FIG. 4. The seasonal depth profile of a phosphate-phosphorus (μg A/L).

that (1) the supply of these nutrients to the euphotic zone is severely limited and (2) they are utilized by the phytoplankton almost as quickly as they are made available. Consequently their concentrations at any one time give only a very rough picture of their availability and rates of utilization by the plants.

Figs. 4-6 show the seasonal distribution of phosphate, nitrite plus nitrate and silicate respectively in the upper 900 m. Silicate analyses were not begun until August, 1958. Phosphate values range from 0·02 to 0·16 μg AP/L, nitrites and nitrates from undetectable amounts to 1·8 μg AN/L, silicates from 0·3 to 1·8 μg ASi/L in the

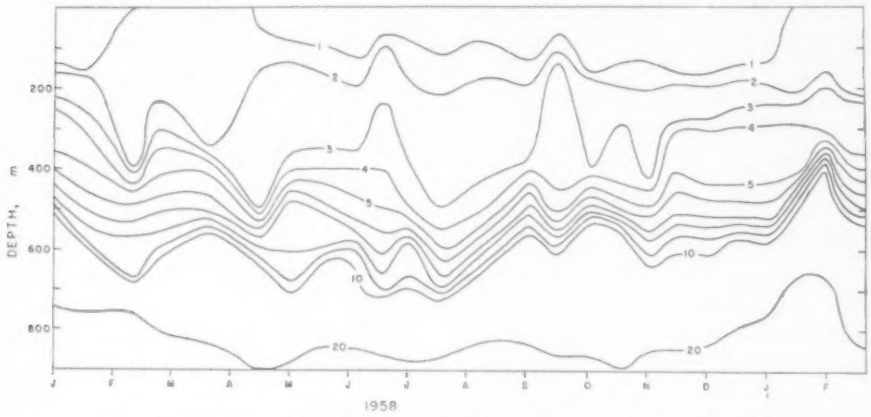

FIG. 5. The seasonal depth profile of nitrite and nitrate-nitrogen (μg A/L).

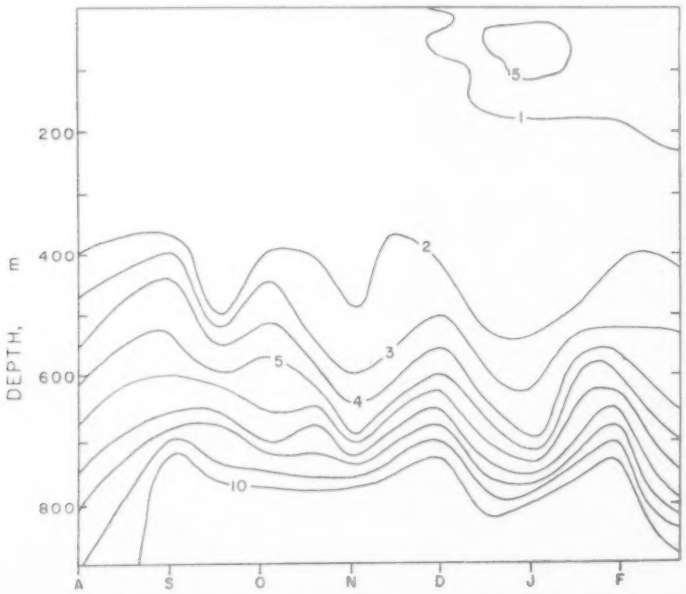

FIG. 6. The seasonal depth profile of silicate-silicon (μg A/L).

upper 100 m respectively. In the case of the nitrogen and phosphorus compounds, the minimum surface values occurred in summer and when the water was stratified, the maxima in the winter and early spring when the greatest vertical mixing occurred. Silicate showed little evidence of a seasonal cycle in these surface waters.

Vol 6 1959

The variations in the nitrogen and phosphorus compounds are obviously correlated with and controlled by the hydrographic cycle described in the previous section. This is a familiar relationship in oceanography resulting from the fact that these materials are utilized nearly to exhaustion in the summer, removed from the euphotic zone through sinking of the phytoplankton, and thereby rendered unavailable for plant growth until winter and spring mixing returns them to the surface. Further discussion of the subject will be deferred until it is considered together with the phytoplankton and its rate of primary production in the final section. However, one important feature may be brought out here. The constancy of the 18° water with respect to its temperature and salinity was emphasized in the preceeding section. The same is very nearly true of its inorganic nitrogen and phosphorus concentrations (2-3 μg AN/L and 0·10–0·15 μg AP/L respectively) despite the fact that they are non-conservative properties. Within and below the permanent thermocline, the concentrations of these elements increase rapidly. The relative constancy of all of these properties is indicative of minor upward transport of nutrients from the region of the permanent thermocline and below. As far as concerns the phytoplankton and the factors which contribute to its growth, the Sargasso Sea may be considered an oligotrophic basin roughly 400 m deep almost out of contact with the rest of the ocean.

Vol.
6
959-60

PHYTOPLANKTON STANDING CROPS

Fig. 7 shows the seasonal profile of chlorophyll *a* concentrations to a depth of 100 m, roughly contoured in the following categories: $<$ 0·10, 0·10–0·25, 0·26–0·50 and 0·51–1·00 mg/m^3. The striking contrast between winter, homogeneous and summer, stratified distribution, is at once obvious. Beginning in the late fall of 1957, we see what appears to be the remnants of a fall flowering in the deeper water. This is followed by moderately high levels (0·35–0·50 mg/m^3) evenly distributed with depth and remarkably constant in concentration which persists throughout the winter and early spring. By early April a spring flowering of short duration appeared, lasting for about one month in the surface but then sinking and remaining in the lower half of the euphotic zone for another month. Peak chlorophyll concentrations in this 'bloom' were about 1·00 mg/m^3. There followed the summer period with thermal stratification of the surface waters and extremely low surface chlorophyll (0·1 mg/m^3) increasing with depth to about 0·5 mg/m^3 at 100 m. As soon as surface cooling and mixing started in September, 1958, the surface chlorophyll picked up slightly and a short-lived, minor flowering occurred in November when the pigments again returned to a constant depth distribution to 100 m. In the early winter of 1958-1959, the pigments were somewhat lower than in 1957-1958, but returned to about the same level by mid-January. A striking difference in the second year was the 'spring' flowering which began in mid-January and lasted until early in March, two and a half months earlier than in 1958, slightly lower in magnitude (maximum of 0·75 mg/m^3 on February 26), and lasting for about twice as long. By the end of March, the phytoplankton were again vertically stratified to 100 m, repeating but two months in advance of the 1958 cycle.

On June 6 and December 18, 1958, chlorophyll measurements were made to 400 m. These data, together with temperatures and inorganic phosphorus values are shown in Fig. 8. On both dates the surface waters were thermally stratified, throughout

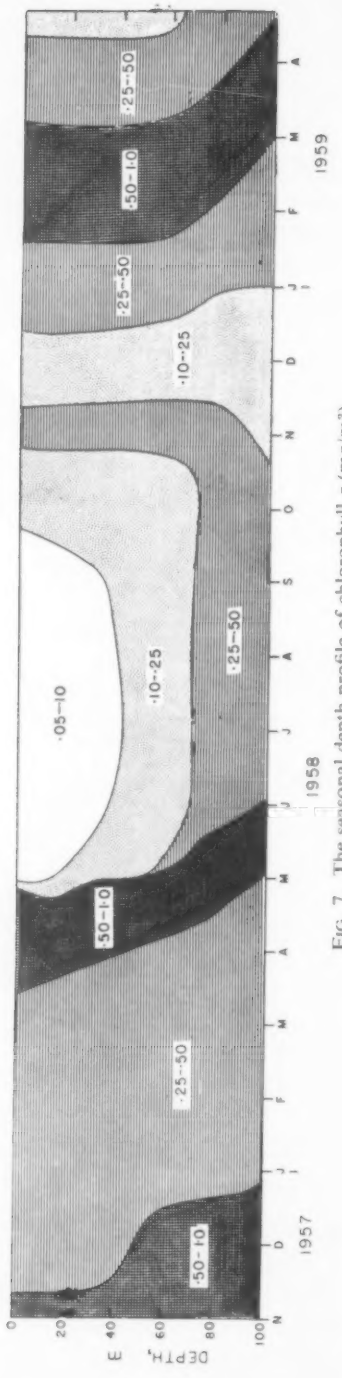

FIG. 7. The seasonal depth profile of chlorophyll *a* (mg/m³).

Vol 6 1959

the upper 100 m in June and from 50-100 m in December. In each case the surface chlorophyll values were low but then increased to maxima at or near 100 m. The nutrients were also low at the surface and began to increase with depth at about 100 m. This is shown by the phosphate distribution and is also true of nitrate, which increased from 0·29 and 0·23 μg AN/L at 50 m to 1·3 μg AN/L at 150 m on the two dates.

Chlorophyll maxima at or near the lower limit of the euphotic zone appear to be a common distributional pattern whenever surface waters are thermally stratified (see Fig. 7). J. H. STEELE and C. S. YENTSCH (unpublished data) have recently considered the causes of this intermediate chlorophyll peak. It appears to be composed

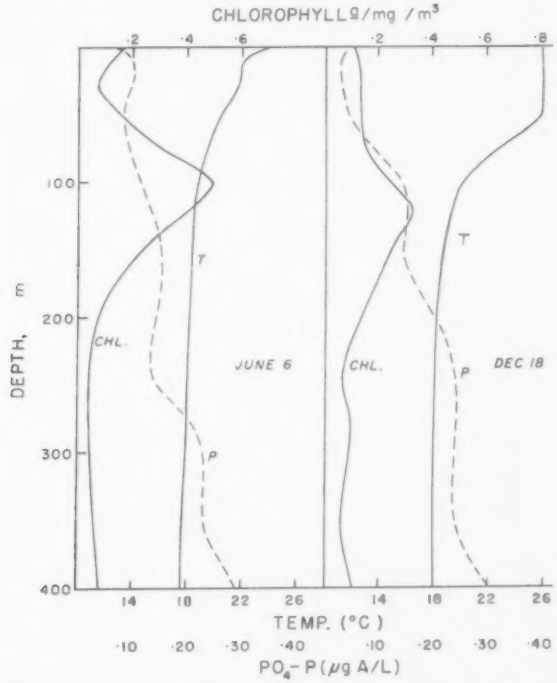

FIG. 8. The vertical distribution of chlorophyll *a* (Chl), temperature (T) and phosphate-phosphorus (P) on June 6 and December 18, 1958.

of the same organisms which inhabit the surface waters and does not represent a distinct population. It therefore probably results from the sinking of organisms from the surface layers. Assuming no differential grazing in the upper 150 m, the only thing which could cause the organisms to accumulate at a depth below that at which optimum photosynthesis occurs (i.e. 25–50 m) would be a change in their sinking rate. STEELE and YENTSCH postulated that nutrient deficient phytoplankton sinking from the surface layers might reduce their sinking rate when they encounter the relatively rich water at the base of the euphotic zone. They substantiated this theory experimentally by showing that the settling rate of nutrient starved diatoms was significantly reduced if the cultures were enriched.

This seems to be a logical explanation for the summer chlorophyll distribution which we have observed. However, phytoplankton counts, which are not yet completed, will be necessary to confirm the fact that the phytoplankton at 100 m, are not, in fact, a separate community, or that the chlorophyll peak at that depth is not simply a matter of 'shade adaptation' in which the chlorophyll per cell is much higher than at the surface. The latter is probably true to some extent, but it seems unlikely that it alone could account for the observed differences in pigment concentrations.

It is noteworthy that some chlorophyll was found all the way to 400 m, though the concentration decreased rapidly from the 100–150 m peak to extremely low concentrations in the deeper waters. We have investigated the problem of whether the chlorophyll at the aphotic depths below 100 m is 'active' in the sense of being capable of carrying out photosynthesis. C^{14} uptake of samples from 0 to 400 m was measured at 1500 ft candles for 24 hrs in a water-cooled, artificially-illuminated incubator. The results, corrected for dark C^{14} uptake, showed no carbon assimilation in any of the samples from below 100 m (Table 3). This may have resulted from light

Table 3. *Photosynthesis* (C^{14} *uptake*) *for* 24 *hrs at* 1500 *ft candles of water from different depths at Station S on December* 18, 1958

Depth in metres	mg *carbon assimilated per* m^3 *per* 24 hr
0	3·05
17	2·10
34	2·20
60	2·90
120	0·47
150	0·00
200	0·00
250	0·00
300	0·00
350	0·00
400	0·00

inhibition of the deep, 'shade adapted' plankton when exposed to the experimental intensity of 1500 ft candles (see RYTHER and MENZEL, in press). Note, for example, that photosynthesis at this constant light intensity decreased with increasing depth of the sample between 0 and 100 m despite the fact that the chlorophyll increased with depth to that level. However, the absorption spectra of pigment extracts from depths below 100 m are quite different from that of plants in the euphotic zone, and it therefore seems likely that the equations of RICHARDS with THOMPSON (1952) are not applicable under such conditions.

PRIMARY PRODUCTION

The values for gross and net primary production in grams carbon assimilated per m^2 per day are shown in Fig. 9. The net production (C^{14} uptake) data for the first four months are rather unreliable for various technical reasons, most important of which being the use of C^{14} solution of insufficiently high activity. Thereafter the data are believed to be accurate within the limits of the methods.

Gross production, based partially on the chlorophyll content of the water, naturally followed closely the chlorophyll distribution. Net production values, with one

exception, were consistently lower, ranging from about 10 per cent to 90 per cent and averaging about 50 per cent of gross production (omitting the first four months). The closest agreement between the two methods occurred at the termination of the spring flowering and throughout the summer of 1958. Beginning in December, 1957, and throughout the early spring of 1958, including the early flowering of that year, a rather constant discrepancy of about 50 per cent between the methods persisted. For the most part, the irregularities in the seasonal cycle of one set of values is closely paralleled by those of the other.

Gross production was relatively high and constant throughout the early winter of 1958, began to increase slowly and irregularly by mid-January, 1959, culminating in a peak of 2·0 g C/m²/day on April 15. Following this maximum, gross production decreased rapidly for the next two weeks, reaching its low, summer level (0.1–0.2 g C/m²/day) by May 1.

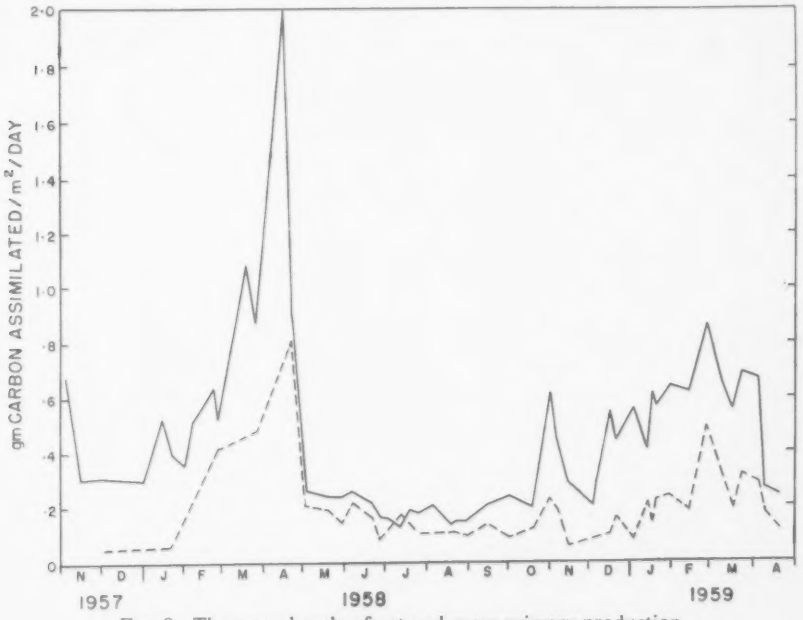

FIG. 9. The annual cycle of net and gross primary production.

In the early part of November there was a brief increase in production which, probably by coincidence, occurred at the same time and attained the same magnitude (0·6 g C/m²/day) as in the fall of 1957. Production then decreased, returning to 0·2 g C/m²/day in December, following which the same gradual and irregular build-up occurred throughout the winter as in 1957-1958, but this time beginning a month sooner and reaching a lower maximum (0·86 g C/m²/day) on February 26. Rather than the sudden drop-off observed in late April, 1958, the decline in the spring of 1959 was also gradual and more irregular. In contrast to the brief flowering of relatively high magnitude in the spring of 1958, the 1959 bloom was more moderate but spread over a much longer period.

Unfortunately, no measurement of net production was made at the peak of the 1958 flowering. Four days later on April 19, when gross production had dropped to 1·5 g C/m²/day, net production was only 0·8, or about half of gross production. Two days, after that, however, and thereafter for several months, the two measurements were in close proximity. The discrepancy on April 19 may have been due to an error of measurement, but the values for March 25 show the same feature, which indicates that net production was only about half of gross production during the build-up of the spring flowering, between 75 per cent and 90 per cent of gross production during the decline of the bloom and the summer period of minimum production. This same trend was observed when production was measured by these methods in the continental shelf waters off New York (RYTHER and YENTSCH, 1958). It may represent some systematic error in either of the methods under conditions of high productivity. One would ordinarily expect that the ratio net/gross production might be highest during periods of most rapid growth of the algae. This is not necessarily so, however. A review by RABINOWITCH (1951) of several investigations shows that respiration is markedly stimulated by the products of photosynthesis. It follows that starved plants respire less, have a greater P/R ratio, and hence a closer relationship between gross and net production than healthy growing cells, which must metabolize at a greater rate to convert their first product of photosynthesis into new cellular materials. Thus the observed relationships between net and gross production are not incompatable with the evidence of plant physiology, though this is clearly an area in which further experimental work is needed.

The annual rate of production obviously depends upon the choice of time within the twelve-month period. For 1958 (January-December) gross and net production were respectively 160 and 72 g C/m². Daily gross production ranged from 0·13 to 2·00 with an annual mean of 0·44 g C/m²/day, which agrees more closely with the much maligned values of RILEY *et al.* (1949) than with any of the more recent estimates. Net production ranged from 0·05 to 0·83 and averaged 0·20 g C/m²/day. The minimum (observed in November) is identical to that reported by STEEMANN NIELSEN for the Sargasso Sea (Table 1) in June. Our June values ranged from 0·09 to 0·17 g C/m²/day, not appreciably different from STEEMANN NIELSEN's, though his values are increased by a respiratory factor to make them equivalent to gross production.

Most of the year's production occurred during the winter build-up and the spring flowering, if indeed the two can be separated. In the four-month period between January 1 and April 30, 1958, gross production was 92 g C/m², almost 60 per cent of the year's total. If we consider the equivalent period of the following year to extend from December 1, 1958 through mid-April, 1959, gross production amounted to 76 g C/m²/year, or 82 per cent of the previous year's value for the similar period. It appears likely, therefore that annual production does not vary greatly from one year to the next, as if the total amount of organic matter which may be produced is more or less fixed in some way, perhaps by the available nutrients. On the other hand, the way in which this potential is expended over the year is highly variable and probably dependent upon local and transient meteorological and hydrographic features which will be discussed in the following, final section.

THE ENVIRONMENTAL CONTROL OF PRIMARY PRODUCTION

It was stated earlier that, as far as the phytoplankton were concerned, the Sargasso

Vol 6 1959

Sea could be considered as a basin 400–500 m deep, its bottom the discontinuity between the 18° water and the permanent thermocline. This obviously is not entirely true, for sinking and downward transport to 1000 m or more by the migratory zooplankton must result in some loss of organic matter from the surface layers. The nutrients bound in this material must be replenished in some way to the upper water if the fertility of the area is not to be progressively diminished. Yet the constancy of the temperature and salinity of the surface layer is a convincing argument against the occurrence of any large-scale mixing between the thermocline and the overlying waters. It follows, then, that the nutrient elements must complete almost their entire biochemical cycle of assimilation and regeneration within the upper 400 m. It also follows that production must be severely limited by the relatively low concentrations of nutrients in this layer.

Within the upper 400 m, the highest concentration of nitrite plus nitrate nitrogen and phosphate phosphorus are about 1200 mg atoms and 60 mg atoms/m² respectively. If all of these nutrients were incorporated into phytoplankton of normal elementary composition, the resulting crop would be equivalent to some 75–100 grams carbon/m², which is roughly the value obtained for net annual production. If the nutrient regenerative processes were slow, one might expect that the levels of inorganic nitrogen and phosphorus would decline at a rate inversely proportioned to primary production. REDFIELD *et al.* (1937), STEELE (1956) and others were able to estimate production rates by the disappearance of phosphorus from the surface waters of temperate latitudes. However, no such simple relationship exists in the Sargasso Sea. The dissolved nutrients in the 400 m water column vary only slightly and in an irregular and unpredictable way. For example, during the winter and spring flowering of 1958 there was a net production of 46 g C/m² between January 1 and May 1. This is equivalent to a utilization of 570 mg AN and 38 mg AP/m², half the amount present. Yet within that period there was a loss of only 250 mg A/m² of nitrogen and a *gain* of 35 mg A/m² of phosphorus. Clearly the regenerative processes occur so rapidly in these semi-tropical waters that the concentrations of inorganic nutrients present at any time are almost without significance.

It may, of course, be argued that these irregularities in the upper 400 m merely reflect exchanges with the deeper, thermocline waters. Unfortunately the limitations of the analytical methods result in absolute errors which are much greater in deeper waters, where concentrations are higher, and probably obscure the relatively small losses and gains which would be necessary to balance the changes in the surface waters. A further complication is the vertical displacement of properties in the thermocline region by as much as 100 m from one set of observations to the next (Figs. 2–6) which SCHROEDER *et al.* (1959) have attributed to short-period internal waves. Because of these, the homogeneous layer should technically be considered with respect to a density discontinuity rather than a fixed depth, but in view of the limitations of the analytical methods the need for such a refinement is hardly indicated.

There is, however, a gross seasonal cycle of nutrients in the upper 400 m which illustrates at least qualitatively the environmental control of production. For example, in the winter and early spring, when water is mixed to depths as great as 400 m, the nutrients reach their highest values at the surface (1–2 μg AN/L, 0·1–0·2 μg AP/L) and their lowest values at the depths 100–400 m. The euphotic zone remains almost constant throughout the year at about 100 m. In winter, the phytoplankton is evenly

distributed from 0–400 m and the entire population, though limited by light, is able to grow. When thermal stratification of the surface waters first begins to develop, the plants are maintained in the euphotic zone at relatively high nutrient concentrations, giving rise to a vigorous but short-lived flowering. A typical example of this is the maximum of production on April 15, 1958 of 2·0 g C/m²/day, which approaches the highest recorded values for primary production anywhere (RYTHER, in press). Since growth at this rate would require an average daily consumption of 0·25 μg A/L of nitrogen and 0·02 μg A/L of phosphorus within the 100 m euphotic zone, it is obvious that nutrient exhaustion must quickly terminate this flowering. The bulk of the spring bloom then sinks and accumulates at 100–150 m (Figs. 7 and 8) and the organisms or their decomposition products remain there until mixing returns them to the surface the following winter. The low level of summer production is undoubtedly dependent upon the rapid partial re-cycling of nutrients *in situ* in the euphotic zone, perhaps augmented by occasional mixing below 100 m during summer storms.

There is nothing unique or even unusual in the general nature of this cycle. What distinguished the Northern Sargasso Sea from more temperate regions are the following characteristics : (1) a relatively shallow winter mixed layer, (2) greater solar radiation in winter, (3) clear water, (4) low level of nutrients in the mixed layer and (5) the rapid re-cycling of nutrients, possibly due to higher temperatures. As a result of this combination of factors, the levels of maximum and minimum production in these semi-tropical waters are less extreme but the annual rate is as high as or higher than that of temperate and northern waters (see, for example, RYTHER and YENTSCH, 1958).

As we have shown, the degree of winter mixing is variable and apparently dependent upon winter air temperatures and perhaps winds. This effects the timing and nature of the seasonal cycle of production, but does not appear to influence the overall annual rate as long as the water becomes mixed to a depth significantly below the euphotic zone. By oceanographic criteria, a separation between semi-tropical conditions must exist very near the latitude of Bermuda. A short distance to the South, surface temperatures never fall below 20° and winter mixing probably never extends below 100–150 m. Under these conditions it is likely that primary production proceeds at a low rather steady rate throughout the year, comparable to summer conditions in Bermuda. The results of this study cannot therefore be taken as ' typical ' of the Sargasso Sea or the open ocean in general, since probably neither is as uniform as is generally implied.

Woods Hole Oceanographic Institution, Woods Hole Massachusetts.

Contribution No. 254 *from Bermuda Biological Station*

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Vol 6 1959

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BOOK REVIEW

Conference on Physical and Chemical Properties of Sea-Water, Easton, Maryland, September 4–5, 1958. National Academy of Sciences, National Research Council, 1959. Publication 600, 202 pp.

The book gives a report of the Conference held in Easton, Maryland 4-5 September, 1958, at which 40 outstanding American scientists met to discuss some of the requirements for observing and interpreting the physical and chemical properties of sea-water.

Three committees were appointed to draft recommendations and suggestions for future work: a Committee on Field Measurements, a Committee on Chemical Properties, and a Committee on Physical Properties.

As a focus for the discussions, 17 papers covering the different fields on chemical and physical oceanography were read. The book first gives the reports from the three committees, and then the 17 papers, each of which is followed by a summary of the discussion.

About one third of the book deals with physical oceanography, discussing: The classical hydrographic procedure, geostrophic currents, adiabatic temperature gradient and thermodynamics of sea-water. The next part deals with chemical oceanography: The chlorinity-density relationship and the problem of the variations of the relative proportions of the major constituents of sea-water, assessment of the accuracy and precision of data, and the possible role of biological processes in the composition of sea-water. A third part describes analytical techniques and various new apparatuses which are, or can be, developed for the determination of the salinity or density of sea-water.

I can recommend this book as being of great value to all working oceanographers. It gives a picture of the present situation stating what is known and unknown in physical and chemical oceanography, and pointing out ways to be followed in the future. It is an incentive to reconsider these problems and to discuss their solution.

The book is printed in rota-print and is unbound. However, good figures and many references make this book attractive and very useful to all who are working in this field.

E. FØYN

Vol
6
1959

BOOKS RECEIVED

Annual Report of the National Oceanographic Council 1958–59. Price 5s. net. Cambridge University Press.

ERRATA

TORBEN WOLFF : The hadal community, an introduction.

p. 95, sec. (5) line 3 : For (43 per cent) read (74 per cent).

p. 100, bottom line : For (49 per cent) of the 127 identified species, read (74 per cent) of the 85 identified species.

p. 101, line 3 : For 6000 m read 6000 m : 127 species.

p. 102, line 6 from bottom : For 11,700 m read 10,700 m.

p. 106, line 19 : For Ascidiacea 1 read Ascidiacea, *Galathea Rep.* **1**.

p. 112. After 163 read 164.

p. 112. 167 : for spärcfli read spärcki.

p. 112. 170 : should not be in bold type.

p. 113. 181 : should not be in bold type.

A. P. LISITZIN and A. V. ZHIVAGO: Marine geological work of the Soviet Antarctic Expedition, 1955—1957. Gummed slip to be attached to Fig. 7, p. 86, DEEP-SEA RESEARCH, Vol. 6, No. 2.

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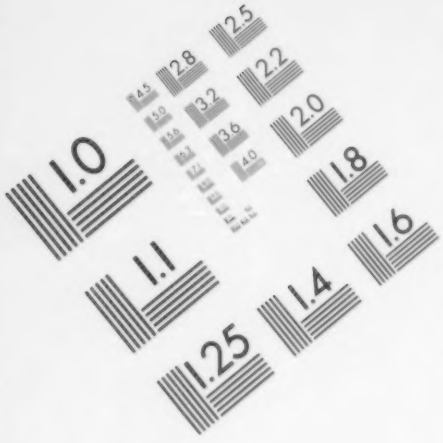

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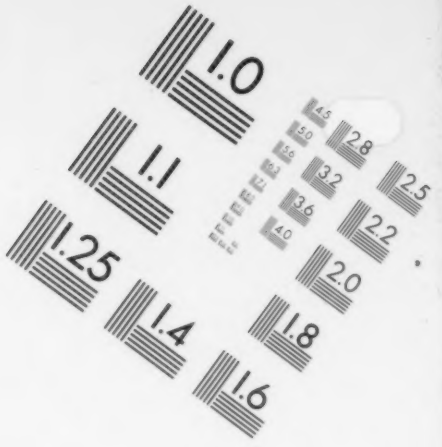

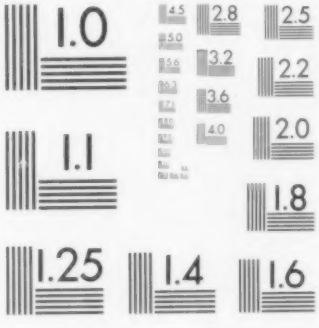

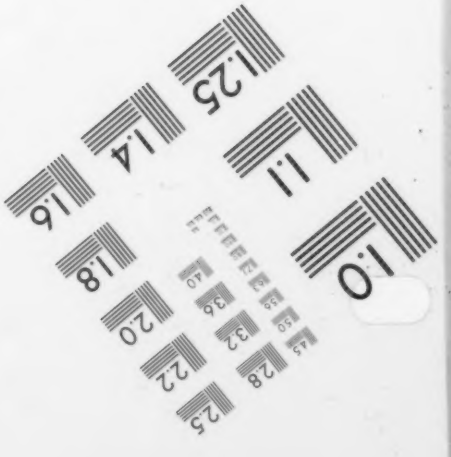

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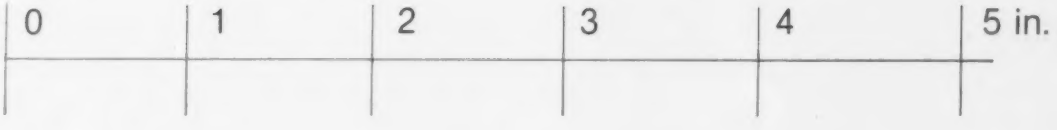

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